Archean and Proterozoic Geology of the Lake Superior Region, U.S.A., 1993

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P.K. Sims *and* L.M.H. Carter, *Editors*

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FOREWORD

This volume was written to update a report on the Precambrian geology of the Lake Superior region prepared for the Geological Society of America's Decade of North American Geology (DNAG) volume (Volume C–2) on the Precambrian–Conterminous U.S., edited by J.C. Reed, Jr., and six others, and published in 1993. The chapter on the Lake Superior region is one of seven comprehensive reports in Volume C–2. The Lake Superior chapter was written in 1985 and page proof was prepared in 1987, 6 years before publication of Volume C–2. Accordingly, much of the chapter is out of date. Sections on the Archean, Early Proterozoic, and Middle Proterozoic Midcontinent rift system are particularly obsolete because of substantial new knowledge resulting from ongoing studies by individuals from several organizations.

Most of the authors of sections in this Professional Paper were authors of sections in the Lake Superior chapter in DNAG Volume C–2. These authors have continued research since the mid-1980's and continue to maintain expertise in their field of specialty. The authors of the section on the Midcontinent rift in this volume, however, were not involved with preparation of the DNAG volume; they have been major investigators in the exciting recent seismic-profiling and petrochemical studies of the rift. For completeness, short summaries of those geologic entities that have received relatively little study since the mid-1980's, such as the Middle Proterozoic Wolf River batholith, are included in this report.

This report is written for the same readers as the DNAG volumes, that is, geologically knowledgeable people, both within North America and abroad, who want an overview of the Precambrian geology of the United States segment of the Superior province and adjacent Early and Middle Proterozoic terranes of the Canadian Shield.

I thank each of the authors for thoughtful preparation and for timely response in completing manuscripts. G.B. Morey and W.F. Cannon ably reviewed the entire report and made numerous helpful suggestions. I am grateful also for the careful, thorough technical editing of co-editor L.M.H. Carter.

P.K. Sims

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P.K. Sims *and* L.M.H. Carter, *Editors*

INTRODUCTION

By P.K. Sims

The Precambrian rocks of the Lake Superior region have been studied nearly continuously for more than a century, mainly because of the early discovery of large, valuable iron deposits in Minnesota, Michigan, and adjacent Wisconsin and copper deposits in the Keweenaw Peninsula, Michigan. Tremendous advances in knowledge of the geology of the region have been made particularly during the past three decades because of available detailed aeromagnetic and gravity anomaly maps that cover much of the region, nearly complete coverage of modern topographic maps, the development of sophisticated isotopic dating techniques, and the application of seismic-reflection profiling to major geologic problems, particularly the Middle Proterozoic Midcontinent rift system. The Precambrian rocks can now be placed confidently in a general stratigraphic framework that places constraints on detailed, more local studies.

The Precambrian rocks of the Lake Superior region lie along the south margin of the Superior province of the Canadian Shield (fig. 1), and underlie large parts of the area of Minnesota, Wisconsin, and Michigan. They are overlapped to the west, south, and east by flat-lying sedimentary rocks of Paleozoic and Mesozoic age of the interior platform and are discontinuously covered by Pleistocene glacial deposits. The present landscape is subdued, and is inherited largely from erosional and depositional landforms of the Pleistocene glaciation. Inliers of Precambrian rocks are exposed locally in the covered areas, mainly in the valleys of major rivers in southern Minnesota and Wisconsin, where erosion has dissected the Phanerozoic rocks.

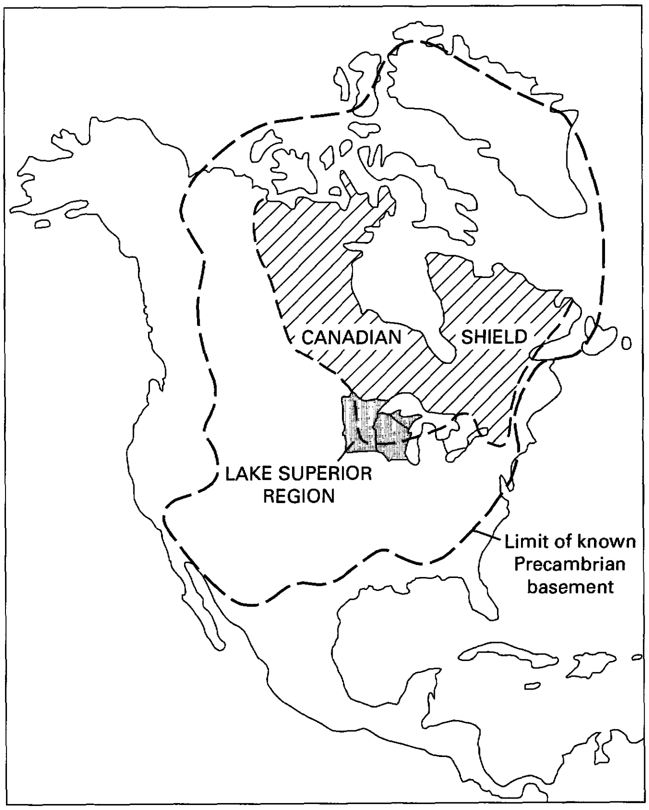

Figure 1. Location of Lake Superior region relative to the Canadian Shield and to the inferred extent of Precambrian rocks of the North American craton. Compiled from various sources.

This report is organized in the same general way as the DNAG report, that is, around a geochronometric framework. Descriptions of the Archean terranes, the Early Proterozoic Penokean orogen, and the Middle Proterozoic intracratonic igneous and sedimentary rocks, including the Midcontinent rift system, are followed by brief overviews of the gravity and magnetic characteristics of the several terranes and the metallogeny of the region. Emphasis is generally placed on new data obtained since about 1985 and new and (or) refined interpretations. Accordingly, the reader should refer to the DNAG chapter as well as this report for an overall view of the Precambrian geology of the Lake Superior region. The final section in this report relates the Precambrian geology of the Lake Superior region to that in adjacent regions.

The term "terrane" is used in the Lake Superior region in the same way as in the North America cordillera (Jones and others, 1977). A terrane constitutes a distinctive crustal segment characterized by unique stratigraphy, structural style, geophysical characteristics, and age, and commonly is fault bounded.

The term "province," as used herein, follows Canadian terminology for Archean rocks (>2,500 Ma; Hoffman, 1989). An Archean province is a contiguous area of continental lithosphere bounded by Proterozoic orogens inferred to be suture zones. "Subprovinces" are defined on the basis of lithologic and tectonic differences within a province, that is, differences in structural trends, ages of orogenic units, rock types, metamorphic grade, and geophysical characteristics, or a combination of these (Card and Ciesielski, 1986; Card, 1990).

GEOLOGIC FRAMEWORK

By P.K. Sims

The Precambrian rocks in the Lake Superior region record more than 2.5 b.y. of geologic time (pl. 1). This interval of geologic time is not continuously recorded in layered and intrusive units, but instead is punctuated by specific rock-forming and tectonic events that can be deciphered from geologic relations and placed in a chronometric framework by isotopic dating.

The Lake Superior segment of the Canadian Shield is a collage consisting mainly of Archean cratonic elements (3.6–2.5 Ga), assigned to the Superior province, a flanking Early Proterozoic (≈2.1–1.84 Ga) orogenic belt (Penokean orogen), and a Middle Proterozoic (≈1.1 Ga) intracratonic rift assemblage (Midcontinent rift system) that transects older Archean and Proterozoic rocks. In addition, intracratonic igneous and sedimentary rocks in the approximate age range 1.82–1.63 Ga overlap older rocks on the south, and anorogenic intrusions (≈1.47 Ga) intrude older rocks in central Wisconsin. The rocks represent several major crust-forming events, reactivation of Archean basement rocks locally within the Penokean orogen, the erosion and deposition of epicratonic successions, and the local intrusion of anorogenic magmas.

The Archean in the Lake Superior region comprises the southern part of the Superior province, a large, relatively undisturbed Archean craton having an exposed area of more than 2 million km^2 (Card and Ciesielski, 1986). In Canada, the Superior province is made up of four rather distinct subprovinces that differ markedly from one another and form a strongly linear regional structure. These subprovinces (Card and Ciesielski, 1986) consist dominantly of (1) plutonic rocks, (2) volcanoplutonic rocks, (3) metasedimentary rocks, and (4) high-grade gneisses. Volcanoplutonic and metasedimentary subprovinces are the predominant rock assemblages in the United States segment of the Superior province (Card, 1990).

Three subprovinces delineated in Canada can be projected into the United States (Southwick, this report). From north to south, these are (1) the Wabigoon subprovince, a volcanoplutonic terrane, (2) the Quetico subprovince, a dominantly metasedimentary terrane, and (3) the Wawa subprovince, a volcanoplutonic terrane. A fourth subprovince, not recognized in Canada—the Minnesota River Valley subprovince, a gneiss terrane—lies to the south of the Wawa subprovince. The gneiss terrane forms the southernmost part of the Superior province; it is partly overlapped to the south by Early Proterozoic rocks of the Penokean orogen and by younger quartzite. High-angle transcurrent faults that formed in a dextral transpressive environment mainly separate the Wabigoon, Quetico, and Wawa subprovinces, but at places the Quetico-Wawa boundary is a metamorphic or lithologic boundary rather than a fault (Southwick, this report). The boundary between the Wawa subprovince and the Minnesota River Valley subprovince is a major structure named the Great Lakes tectonic zone (GLTZ; Sims and others, 1980). In Minnesota, the GLTZ, locally named the Morris fault, has been interpreted from geophysical data as a north-dipping, south-verging structure (Gibbs and others, 1984); in northern Michigan, however, the GLTZ has been interpreted from outcrop data as a north-verging structure (Sims, 1991a; Sims, 1993; Sims and Day, 1992). This dichotomy in interpretation is discussed in a following section.

In previous reports on the Lake Superior region, summarized in the DNAG volume (Sims and others, 1993), the rocks comprising the three Archean subprovinces north of the Great Lakes tectonic zone were collectively called the Archean greenstone-granite terrane, and the Minnesota River Valley subprovince, to the south, was called the Archean gneiss terrane. These two designations were used to distinguish predominantly juvenile volcanic and plutonic rocks of Late Archean age to the north from mainly older gneisses of uncertain origin to the south (Sims and others, 1980). In previous reports, the gneisses of the Minnesota River Valley subprovince were not considered to be a part of the Superior province.

The Early Proterozoic rocks in the Lake Superior region mainly comprise the Penokean orogen, a major orogenic belt that extends from the Lake Superior region northeastward through the Sudbury basin to the Grenville front (Sims and others, 1980), and is inferred to extend in the subsurface southwestward through southern Minnesota and northwestern Iowa into Nebraska, where it is truncated by the younger (1.8–1.63 Ga) Early Proterozoic Central Plains orogen (fig. 2; Sims and Peterman, 1986; Sims, 1990d; Sims, Peterman, and others, 1991). The Penokean orogen consists of a belt of island arc volcanic and plutonic rocks (1.9–1.84 Ga; Wisconsin magmatic terranes), a maximum of 200 km wide (Sims and others, 1989), juxtaposed with an asymmetrical continental margin assemblage on the north. The continental margin assemblage consists of older passive margin sedimentary and volcanic rocks overstepped northward by an asymmetrical foredeep assemblage, both of which were deposited on an Archean basement. The structure along which the two terranes are juxtaposed, the Niagara fault zone in Wisconsin and Michigan and the Malmo discontinuity in Minnesota (Southwick and Morey, 1991), is a north-verging paleosuture. The continental margin assemblage includes the world-class Lake Superior-type iron ores (Morey, 1983b). The arc rocks, which contain several valuable Zn-Cu massive sulfide deposits (Sims and others, 1987 and references therein), were accreted to the continental margin at ≈1.85 Ga.

The intracratonic igneous and sedimentary rocks (unit Xrg, pl. 1) postdate the consolidation of the Superior Archean craton and the Early Proterozoic Penokean orogen, to form the south margin of the vast composite North American craton called "Laurentia" (Hoffman, 1988). Following mild deformation and metamorphism of these rocks at about 1.625 Ga, anorogenic plutonic rocks were emplaced; the largest of these plutons, the Wolf River batholith, is ≈1.47 Ga. The plutonic rocks are components of the 1.5–1.4 Ga Transcontinental Proterozoic anorogenic province of North America (Anderson, 1983).

The youngest major terrane in the region, the Middle Proterozoic (Keweenawan Supergroup; ≈1.1 Ga) Midcontinent rift system, is an intracontinental assemblage of igneous and sedimentary rocks that developed in a rift that aborted before significant crustal separation was achieved. The rocks are dominantly bimodal basalt and rhyolite, which occupy a central horst; gabbro-anorthosite complexes, which were intruded along the Archean–Early Proterozoic unconformity along the margins of the rift; and mainly red bed, but locally carbon-rich, sedimentary rocks.

Since the late Middle Proterozoic, the entire region has been stable and, except for the effects of loading by Pleistocene glaciers, apparently has been in isostatic equilibrium.

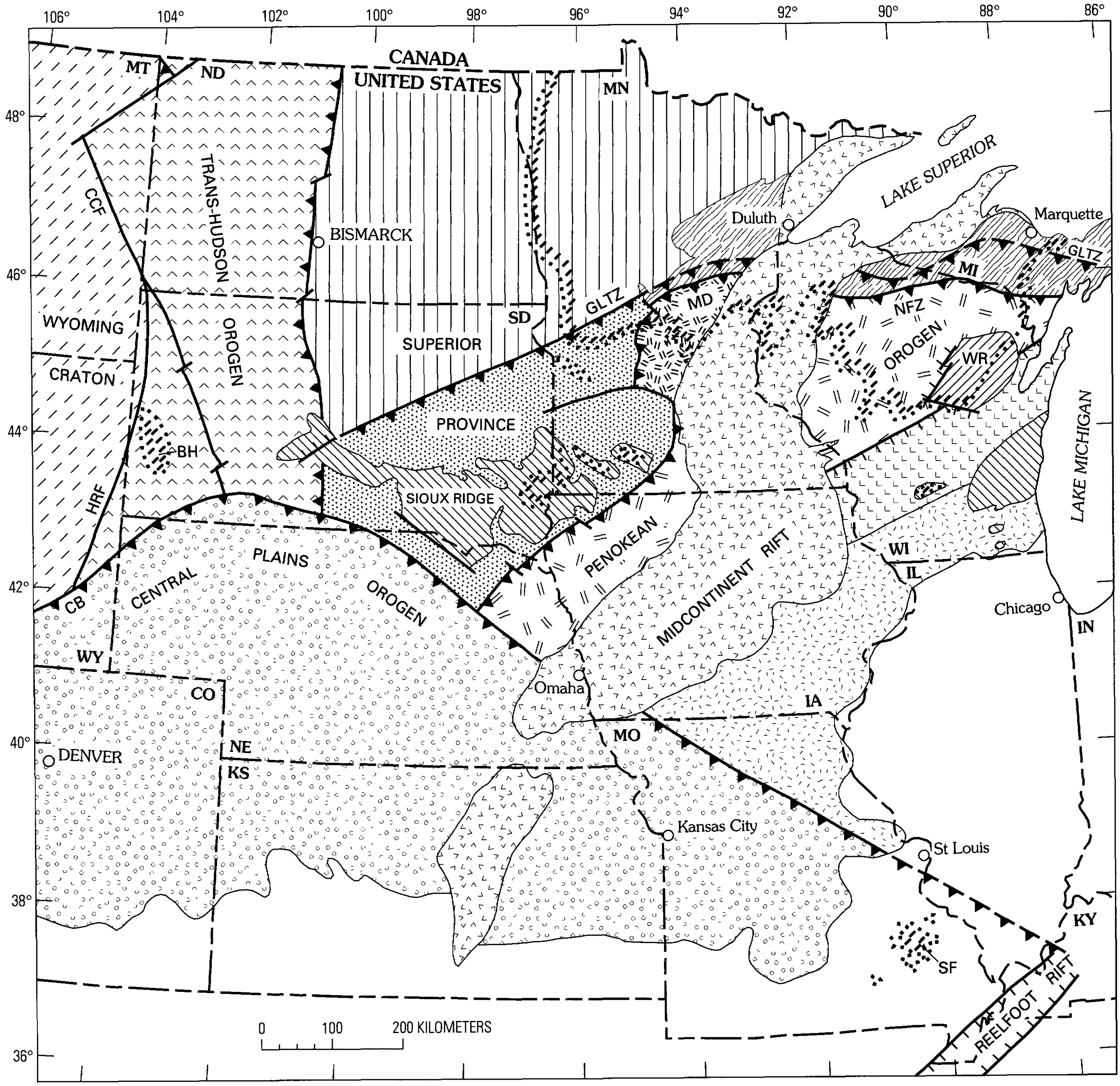

Figure 2 (above and facing page). Simplified tectonic map of Precambrian basement rocks, north-central United States. BH, Black Hills uplift; SF, Saint Francois Mountains; WR, Wolf River batholith.

EXPLANATION

MIDDLE PROTEROZOIC (900–1,600 Ma)

- **Midcontinent rift system (≈1.1 Ga)**
- **Rhyolite and anorogenic granite (1.35–1.48 Ga)**
- **Anorogenic granite and anorthosite (≈1.48 Ga)**

EARLY PROTEROZOIC (1,600–2,500 Ma)

- **Quartzite**—Stipple outlines outcrop area
- **Metamorphic and granitoid rocks of Central Plains orogen (1.6–1.8 Ga)**
- **Rhyolite and anorogenic granite (≈1.76 Ga)**
- **Granite and associated rocks of uncertain age; includes anorogenic granite (1.45–1.50 Ga)**
- **Granitoid and metamorphic rocks (includes Archean gneiss)**
- **Wisconsin magmatic terranes of Penokean orogen (1.35–1.9 Ga)**
- **Sedimentary and volcanic rocks of Penokean continental margin (1.9–2.1 Ga); includes Archean basement rocks**
- **Metamorphic and granitoid rocks of Trans-Hudson orogen (1.81–1.9 Ga); includes scattered Archean rocks**

ARCHEAN (2,500 Ma and older)

- **Volcanic, sedimentary, and granitoid rocks of Superior province (2.6–2.75 Ga)**
- **Gneiss of Superior province (Minnesota River Valley subprovince) (2.6–3.6 Ga)**—Includes local Early Proterozoic rocks
- **Gneiss of Wyoming craton (2.5–3.4 Ga)**

Edge of Phanerozoic rocks

Generalized contact between major rock groups

High-angle fault

High-angle fault—Hachures on downthrown side

Thrust fault—Sawteeth on upthrown side; dashed where covered by Proterozoic rocks

CB	**Cheyenne belt**
CCF	**Cedar Creek fault**
GLTZ	**Great Lakes tectonic zone**
HRF	**Hartville-Rawhide fault**
MD	**Malmo discontinuity**
NFZ	**Niagara fault zone**

ARCHEAN SUPERIOR PROVINCE

By D.L. Southwick[1]

Since late 1985, when the Archean geology of the Lake Superior region was last summarized (Sims and others, 1993), much new work has been done and some important new insights have emerged. The principal contributions have involved field-based structural studies and mapping in Minnesota, Wisconsin, and Michigan, and the development of a unifying model for Archean tectonics, based mainly on work in Canada (Card, 1990), into which the various United States studies appear to fit. In this report the unifying model will be introduced first, followed by brief discussions of recent advances derived from studies in Archean terranes that lie to the north of the Great Lakes tectonic zone in the United States segment of the Lake Superior region.

THE SUBPROVINCE CONCEPT

BACKGROUND

Researchers have known for decades that the Superior province of the Canadian Shield consists of subdivisions that differ from one another in structural style, predominant rock type, metamorphic grade, and temporal history as revealed through geochronology. The origin and tectonic significance of the subdivisions and their boundary zones are not completely understood, but it is beginning to appear that the subdivisions, now generally called subprovinces, were smaller tectonic entities such as volcanic arcs, back-arc basins, accretionary prisms, and seamount chains that were swept together by subduction and related processes of convergent tectonism in Late Archean time (Card, 1990; Thurston and Chivers, 1990). The result was a continental mass of which the Superior province as a whole is a surviving major fragment (Card, 1990; Hoffman, 1988).

The terminology of the subprovinces of the Superior province has evolved along a tortuous path that need not be retraced here. The terminology proposed first by Card and Ciesielski (1986), and amplified by Card (1990), is now widely accepted, and provides a conceptual framework for discussing the Archean geology of the Lake Superior segment of the Superior province.

CHARACTERISTICS OF THE SUBPROVINCE TYPES

The four subprovince types within the Superior province are distinguished fundamentally on the basis of the predominant rock types or rock-type associations they contain, as follows (Card, 1990): (1) *plutonic*, in which intrusive granitoid rocks account for an overwhelming fraction of the terrane; (2) *volcanoplutonic* or ("greenstone-granite"), which contain voluminous sequences of broadly arclike volcanic rocks at low to moderate metamorphic grade, typically in the greenschist or lower amphibolite facies, and extensive suites of intrusive granitoid plutons that are chiefly of "I-type" geochemistry; (3) *metasedimentary*, which contain thick sedimentary sequences, most typically of turbiditic affinity, together with migmatite and paragneiss derived from a metasedimentary protolith by partial melting and (or) injection of granitoid leucosome; (4) *high-grade gneiss*, characterized by extensive tracts of upper amphibolite- and granulite-facies gneisses and commonly geochronologically complex.

Subprovince boundaries in many cases are marked by major fault zones, ductile shear zones, and (or) abrupt metamorphic gradients. However, some subprovince boundaries are essentially conformable stratigraphic contacts between contrasting rock suites (Kehlenbeck, 1976; Borradaile and Spark, 1991) without significant structural dislocation or metamorphic discontinuity along them.

GENERAL ATTRIBUTES OF SUBPROVINCES IN THE LAKE SUPERIOR REGION

From north to south, the four Archean subprovinces in the Lake Superior region are as follows (pl. 1):

1. The Wabigoon subprovince, a volcanoplutonic entity that underlies most of northwestern Minnesota.

2. The Quetico subprovince, a metasedimentary belt that extends westward from the Ontario-Minnesota border lakes between Rainy and Basswood Lakes to a poorly defined termination in the subsurface of west-central Minnesota or eastern North Dakota.

3. The Wawa subprovince (Wawa-Shebandowan subprovince of several Minnesota Geological Survey reports), another volcanoplutonic entity that includes the Vermilion district (Sims and Southwick, 1985), the Virginia "horn" (fig. 3), the Giants Range batholith, the greenstone belt of northern Itasca County (Jirsa and Boerboom, 1990), and a broad swath of greenstone belt rocks farther to the southwest in Minnesota. The Ishpeming greenstone belt in northern Michigan (pl. 1) also is assigned to this subprovince. The geophysically defined "quiet zone" that extends across central Minnesota is provisionally included in the

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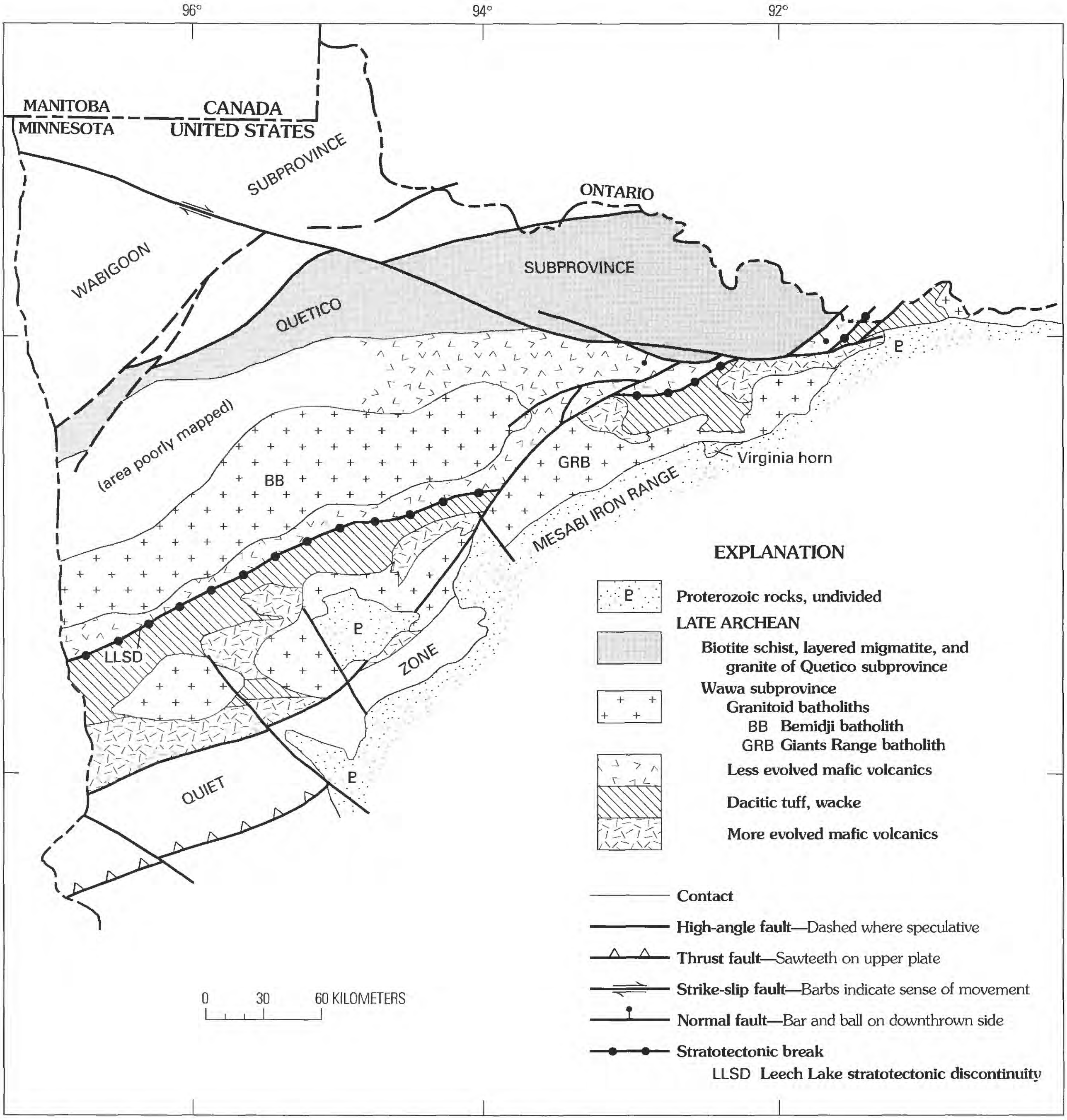


Figure 3. Sketch map of northern Minnesota showing the major geologic units within the Wawa subprovince of the Archean Superior province. Modified from Jirsa and others (1991). Figure 4 gives further detail of faults and disconformities. The interpretation of vergence of Great Lakes tectonic zone differs from that on plates 1 and 2 and in figures 11 and 17.

Wawa subprovince on the basis of sparse data from drilling (Chandler and Southwick, 1990).

4. The Minnesota River Valley subprovince, a high-grade gneiss terrane that underlies most of southwestern Minnesota and is present in variably reworked form within and beneath tectonic elements of the Early Proterozoic Penokean orogen in Upper (northern) Michigan, northwestern Wisconsin, and east-central Minnesota (pl. 1).

Formerly the rock-type association of the first three of these named subprovinces was lumped together as the greenstone-granite terrane (Morey and Sims, 1976; many later publications by others). The rock-type association of the fourth subprovince was lumped as the Archean gneiss terrane, the Minnesota River Valley gneiss terrane, or simply the gneiss terrane to emphasize its contrasting metamorphic grade and other distinctive characteristics. The present authors recommend here that this older usage be discontinued in favor of the subprovince terminology of Card (1990) outlined previously and applied as follows.

The Wabigoon subprovince in northwestern Minnesota contains a wide variety of metavolcanic and metasedimentary rock types that collectively constitute a "greenstone-belt" lithic association (Southwick and others, 1993). Granitoid intrusions of several types also are present (Day and others, 1991). Wabigoon rocks crop out only along the south shore of Rainy Lake, immediately to the east of International Falls (Day, 1990), and in widely scattered exposures southeast of Lake of the Woods (pl. 1). Therefore, geologic mapping and interpretation depend mainly on geophysical data.

Rocks of the Quetico subprovince include the sequences of biotite schist exposed on the Kabetogama Peninsula of Rainy Lake (Tabor and Hudleston, 1991) and north of the Vermilion fault in Lake Vermilion and vicinity near Ely (Bauer, 1985a; pl. 1), various migmatitic and intrusive rocks of the Vermilion Granitic Complex (Southwick and Sims, 1980; Southwick, 1991b; Bauer and others, 1992), and a number of small granitoid stocks (Bauer, 1985b). These rocks pass beneath glacial cover a few kilometers west of long 93° W., and to the west from there they are mapped on the basis of geophysical signature and scattered drilling.

The Wawa subprovince contains the Vermilion district of northeastern Minnesota, near Ely, which is the best exposed and most complete greenstone belt in the United States segment of the Lake Superior region (Sims and Southwick, 1985). The weakly metamorphosed, complexly deformed volcanic and sedimentary rocks of the Vermilion district continue westward along strike into similar sequences in northern Itasca County (Jirsa, 1990; Jirsa and Boerboom, 1990; Jirsa and others, 1992; Southwick, 1991a); from there southwestward they gradually become covered by glacial deposits but can be traced geophysically (Jirsa and others, 1992). The Giants Range batholith (fig. 3) lies between the Vermilion district–Itasca County greenstone belt on the north and another belt on the south that encompasses the Virginia "horn" and sporadic greenstone occurrences along strike to the west. This southern greenstone belt is overlapped unconformably by basal units of the Animikie Group (Early Proterozoic) along the Mesabi iron range, and is covered completely by glacial deposits west of long 93°15' W. Other greenstone belts and plutonic masses (such as the Bemidji batholith) are recognized geophysically in the covered extension of the Wawa subprovince in west-central Minnesota.

The relatively small outcrop area in Upper Michigan (pl. 1), known as the Ishpeming greenstone belt, is isolated by surrounding younger rocks and cannot be traced directly into any of the Archean subprovinces in Canada (Johnson and Bornhorst, 1991). Nevertheless, it is not far along strike from the type area for the Wawa subprovince in Ontario, and its rock types and structural style are consistent with those in the Wawa area. Therefore, the Ishpeming greenstone belt is interpreted as an inlier of Wawa subprovince rocks in Proterozoic terranes.

The Minnesota River Valley subprovince is best known from the petrologic and geochronologic studies of Lund (1956); Goldich, Hedge, and others (1980); Goldich, Wooden, and others (1980); Wooden and others (1980); and Goldich and Wooden (1980) on rocks exposed in the trunk valley of the Minnesota River. In general terms, the subprovince contains gneisses that are predominantly of quartzofeldspathic composition and of upper amphibolite- to granulite-facies metamorphic grade (Perkins and Chippera, 1985). They have undergone a long and complex crustal history that involved events as old as 3,500–3,600 Ma; the subprovince therefore is inferred to have been a preexisting continental fragment or microcontinent in the Late Archean (approximately 2,700 Ma) when it and more juvenile elements were tectonically amalgamated.

Gneissic rocks similar to those in southwestern Minnesota crop out in northwestern Wisconsin and in the cores of domal structures in Upper Michigan (Sims, 1992; also pl. 1). Despite detailed differences in geochronologic evolution, these occurrences are logically grouped with the Minnesota River Valley rocks on the basis of compositional and metamorphic similarity, the presence in both of isotopic components greater in age than about 2,700 Ma, and their similar geographic position with respect to rocks of the Wawa subprovince.

SUBPROVINCE BOUNDARIES

The boundary of the Wabigoon subprovince with the Quetico subprovince is a zone of major strike-slip faulting in Minnesota (pl. 1; figs. 3, 4). The Rainy Lake–Seine River fault to the east (exposed east of International Falls and in Canada) and the Fourtown fault to the southwest (inferred from geophysics and drilling) are interpreted as brittle-ductile, dextral fault zones that were colinear prior to dextral offset by the somewhat younger and more brittle Vermilion fault.

Similarly, the boundary of the Quetico subprovince with the Wawa subprovince is marked by major faults along much of its length. The Haley, Vermilion, and Burntside Lake faults mark the boundary in the Vermilion district (Sims and Southwick, 1985). A few kilometers west of the Vermilion district, however, the subprovince boundary is not a fault contact (Jirsa and Boerboom, 1990); instead it is a conformable or quasi-conformable stratigraphic contact between a volcanic-dominated rock sequence to the south (in the Wawa) and a metasedimentary-dominated rock sequence to the north (in the Quetico).

Assuming that the aeromagnetically defined "quiet zone" is simply a belt of Wawa rocks that has been

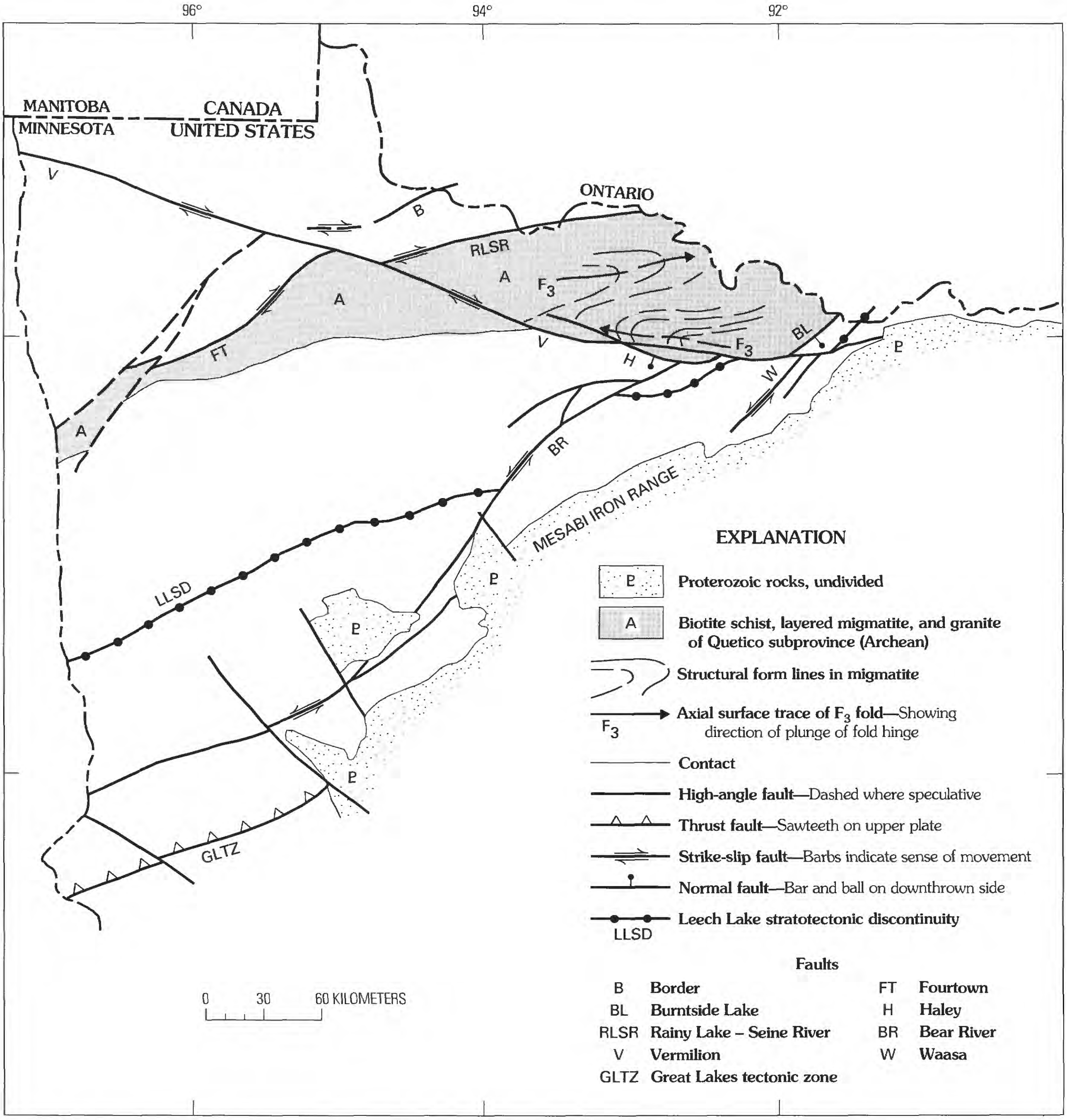


Figure 4. Sketch map of northern Minnesota showing the main faults within the Superior province and the form of largest F_3 folds in the Quetico subprovince. The Leech Lake stratotectonic discontinuity is a major break within the Wawa subprovince that is both a stratigraphic boundary and a zone of faulting.

demagnetized by secondary alteration processes, as is suggested by the results of scattered drilling (Southwick, Meyer, and Mills, 1986; Southwick and others, 1990), the southern boundary of the Wawa subprovince against the Minnesota River Valley subprovince is the Great Lakes tectonic zone (Morey and Sims, 1976; Sims and others, 1980; Gibbs and others, 1984; Sims, 1991a). This structure is enigmatic in Minnesota because it is deeply buried by Quaternary glacial deposits and therefore is not available for direct study. However, it crops out locally in northeastern Michigan (Sims, 1991a) as a zone of ductile shear between granitoid rocks (to the north) and gneisses. The prevailing view is that the Great Lakes tectonic zone (GLTZ) is a major shear zone hundreds of kilometers in length that was established during Late Archean amalgamation of the Superior province, and is a tectonic suture (Sims, 1991a). Later, perhaps at several different times, it was reactivated in response to tectonic events (Sims and others, 1980).

SUMMARY OF NEW WORK IN THE SUBPROVINCES TO THE NORTH OF THE GREAT LAKES TECTONIC ZONE

WABIGOON SUBPROVINCE

New mapping and associated studies by Day (1990) and Day and others (1990; 1991), supported by regional geophysical investigations (Chandler and others, 1987; Bracken and others, 1989a, b), have clarified our understanding of the prominent strike-slip faults in northwestern Minnesota, as portrayed in figure 4. The formerly aligned Fourtown fault and Rainy Lake–Seine River fault have been offset about 30 km dextrally by the northwest-trending Vermilion fault, which clearly indicates that displacement on the Vermilion outlasted displacement on the FT–RLSR system. That the Vermilion fault is oblique to the regional east-northeast trend of major lithologic contacts in northwestern Minnesota is also clear. This oblique relationship strengthens the general argument, developed more fully in a following section, that the Vermilion fault is not itself a fundamental boundary between tectonic terranes.

Day (1990) has verified the position of the RLSR fault zone east of International Falls on the south shore of Rainy Lake, and has identified the presence of several parallel and cross-linking fault strands within it. The east-northeast-trending RLSR fault is the south margin of a dextral-slip wrench zone that has the form of an east-tapering wedge in map view (Poulsen, 1986; also pl. 1). The east-striking Quetico fault is the north margin of the wrench zone; it and the RLSR fault merge a short distance east of the Seine River in Ontario (east of Rainy Lake), and from there eastward a single right-lateral shear zone defines the Wabigoon-Quetico subprovince boundary (pl. 1). To the west of Rainy Lake in Minnesota, the RLSR fault continues as a strong and prominent structure, whereas the Quetico fault weakens and appears to die out. It is probable that motion on the Quetico fault was transferred onto the Border fault (B, fig. 4), another large dextral fault recognized by Day and others (1991) that is subparallel to the RLSR fault and about 25 km northwest of it (fig. 4). Because of its strength and persistence, the FT–RLSR fault system is interpreted to be the Wabigoon-Quetico subprovince boundary southwest of the Rainy River.

In the Rainy Lake area, frictional heat from late movements on the RLSR fault and subsidiary faults in the wrench zone north of it produced globules and veinlets of pseudotachylite that yield Rb-Sr isochron ages of 1,947±23 Ma (Peterman and Day, 1989). In the same area, diabase dikes of the Kenora-Kabetogama dike swarm (Rb-Sr isochron age 2,120 Ma as reported by Warren Beck (reported in Southwick and Day, 1983)) pass directly across the RLSR fault zone with little or no horizontal offset. Thus, the Early Proterozoic shear on the RLSR fault system that is recorded in the 1,947 Ma pseudotachylite ages probably had a very small horizontal component.

QUETICO SUBPROVINCE

Several new studies have focused on deciphering the structural style and structural sequence at and near the margins of the Quetico subprovince, and on ascertaining the relationships of the near-margin structures to structures in adjacent volcanoplutonic subprovinces (Bauer, 1985a, 1986; Tabor and Hudleston, 1991; Bauer and Bidwell, 1990; Bauer and others, 1992). Other studies have dealt with aspects of granite petrogenesis and the formation of migmatites (Day and Weiblen, 1986; Southwick, 1991b). All of this work in Minnesota has benefited from simultaneous studies in contiguous areas to the east in Ontario (Percival, 1986, 1989; Percival and Sullivan, 1988; Percival and Williams, 1989; Borradaile and Spark, 1991).

At and near both margins of the Quetico subprovince, the structural sequence involved first the development of three generations of folds under relatively deep seated conditions, followed somewhat later by ductile to brittle shearing during which the prominent bounding fault systems were formed. The earliest folds (F_1) were tight and quite probably recumbent; they formed in turbiditic rock sequences under metamorphic conditions of the amphibolite facies. These early folds, unlike early folds in the adjacent Wawa subprovince (see next section), developed a pronounced axial-plane schistosity that is parallel to bedding over wide areas and has been reoriented by younger folding events. Folding of the second generation (F_2) was accompanied by widespread emplacement of granitoid rocks, mostly of leucogranite, granodiorite, and trondhjemite composition, and this folding apparently continued past the peak of this igneous activity. F_2 folds are isoclinal to tight, and enfold granitoid layers, lenses, and stringers of diverse size and shape. An associated S or LS fabric is pervasively but variably developed in these folds. F_2 and F_1 folds are nearly coaxial over a wide area, and their long limbs and schistosities together impart a strong east-trending structural grain or fabric to the western part of the subprovince that has been reoriented into large F_3 folds (fig. 4).

The third phase of folding (F_3) was closely associated in time and space with voluminous emplacement of biotite-bearing monzogranite (Late Archean Lac La Croix Granite of Southwick and Sims, 1980). Regional-scale F_3 folds are the most prominent structures in the granitoid-dominated interior of the Quetico subprovince. They tend to be noncylindrical and to lack an associated fabric in roughly the south half of the belt, whereas they tend to be more cylindrical in form and to possess a local L or LS fabric in the north half.

Several lines of evidence indicate that deformation at the immediate margins of the Quetico subprovince and within the neighboring volcanoplutonic subprovinces was transpressional; it involved a component of north-south shortening and a component of dextral shear. Furthermore, evidence is mounting from a number of studies in Canada and the United States to support the conclusion that most of the Superior province was assembled in a regime of dextral transpression that operated over a long span of time (summary in Card, 1990). The situation at the Quetico boundaries therefore fits the overall regional pattern.

However, despite the fact that it probably was deformed in a regional transpressional regime, the interior of the Quetico subprovince displays only slight and local evidence of dextral shear. As a whole, the migmatitic interior lacks the prominent zones of concentrated ductile shear strain and the brittle to semibrittle faults that are so characteristic of the boundaries. The fold geometry and sequence in the interior can be accounted for by repeated episodes of essentially pure north-south shortening. If the interior of the Quetico was in fact affected by dextral shear, the shear strain may have been broadly and subtly distributed across the subprovince because of a lack of competency contrasts and the absence of relatively weak rock units in which shear might be concentrated. This lack of mechanical contrasts on a regional scale probably is due to the abundance of granitoid rock in the core zone. Granitic magmatism would have contributed to isotropic softening of the terrane at the stage when melts were present in the deforming mass, which approximately coincided with the period when F_2 and F_3 folds developed, and the same magmatism later would have contributed to welding and isotropic hardening, after crystallization and cooling occurred. After regional welding, it is likely that continuing shear strain was partitioned into mechanically softer zones of metasedimentary rock near the subprovince margins that were not abundantly injected by granite.

The style and sequence of folding are effectively the same on either side of the Quetico-Wabigoon boundary, and the pattern of strain on either side is consistent with dextral transpression. However, geochronologic evidence indicates that deformation was not strictly contemporaneous in the adjoining terranes. U-Pb zircon data constrain the principal period of folding and fabric formation in the Wabigoon to the interval 2,710–2,700 Ma (Davis and others, 1988, 1989) and also constrain a younger event, dominated by dextral shear, to the interval 2,690–2,680 Ma. In the Quetico, the principal folding event (D_2) must have occurred after about 2,687 Ma, the age of a granodiorite sill deformed in that event (Percival and Sullivan, 1988). The D_3 folding event in the Quetico is associated with granite emplacement that is bracketed between 2,670 and about 2,654 Ma, and therefore it was much younger than the late shearing event recorded in the Wabigoon. Given its faulted condition and the age disparities across it, the contact between the Quetico and Wabigoon subprovinces may be interpreted reasonably as a stratotectonic terrane boundary.

The situation at the Quetico-Wawa boundary appears to be different from that at the Quetico-Wabigoon boundary. Long stretches of this subprovince join are a conformable or moderately unconformable stratigraphic contact between sedimentary and volcanic sequences, without significant fault disruption or abrupt metamorphic contrast (Jirsa and others, 1991; Borradaile and Spark, 1991). The style and sequence of structures are essentially the same across the boundary zone, and the pattern of finite strain is consistent with a history of dextral transpression. Even where the boundary is complexly faulted, as in the Vermilion district in Minnesota (Sims, 1976), strong correlations can be made in the structural sequence developed in the abutting Quetico and Wawa terranes (Bauer, 1985a; Bauer and Bidwell, 1990). Differences in structural style and fabric development are ascribable to differences in the depth and metamorphic environment under which deformation occurred on opposite sides of the faulted boundary zone (deeper to the north, in the Quetico). The Quetico and Wawa subprovinces thus appear to have been a coherent entity through much of their deformational history. Faulting in and near the subprovince join is interpreted as a late manifestation of transpressional shear strain that was concentrated in narrow zones of high competency contrast between sequences of volcanic and sedimentary rock. The Quetico-Wawa subprovince boundary is not a fundamental stratotectonic terrane boundary (Bauer and others, 1992).

WAWA SUBPROVINCE

Recent work in the Wawa subprovince in Minnesota has led to the conclusion that dextral transpression was the mechanism responsible for the observed style and sequence of structures formed in the main period of deformation and fabric development (Hudleston and others, 1988; Bauer and Bidwell, 1990; Bauer and others, 1992). Folds produced during D_2 are tight to isoclinal, and have steep, northeast-striking axial surfaces and hinges that plunge variably but systematically to the east or west. The F_2 folds display a prominent axial-plane cleavage (S_2) and lineation (L_2); these vary regionally in their relative development. The observable strain during this complex D_2 event was strongly partitioned into east-trending zones where the strain was largely constrictional and other, parallel zones where the strain was mainly flattening (Schultz-Ela and Hudleston, 1991). To produce the observed pattern of variation in the symmetry of strain, and the observed fluctuation in maximum stretch from west to east plunges, requires deformation under conditions of noncoaxial shear. The required shear was concentrated in zones of weak, thin-bedded metasedimentary strata that are interdigitated with stronger, less deformable sequences of relatively massive metavolcanic rock. The fault zones at the north margin of the subprovince (discussed briefly in the preceding section) are interpreted as a late manifestation of this overall deformation plan.

A nagging problem in the Vermilion district (northern Wawa subprovince) has been the nature of the F_1 folding. Although F_1 folds have long been recognized (Hooper and Ojakangas, 1971; Sims, 1976; Hudleston, 1976; Ojakangas and others, 1978; Sims and Southwick, 1980, 1985; Hudleston and others, 1988), their size, morphology, orientation, and tectonic significance have been subjects of continuing uncertainty and debate. Substantial evidence now exists to support the following inferences about the F_1 event: (1) The larger F_1 structures have dimensions on the scale of kilometers. This is indicated by the large areas of the western Vermilion district within which easterly plunging F_2 folds face downward and westward in the S_2 cleavage (Jirsa and others, 1992), which requires that the stratigraphic sequence was inverted over large areas prior to F_2 folding. Such broad areas of inverted stratigraphy imply the existence of F_1 fold nappes and extensive preservation of their inverted lower limbs. (2) Outcrop-scale F_1 folds occur chiefly in sequences of thin-bedded sedimentary rocks (iron-formations; medial to distal turbidite sequences). Their morphology comprises tight to isoclinal folds of diverse shapes and styles that are suggestive of soft-sediment origin, sheath folds, and zones of complex fold interference. Axial-planar fabric is weak or absent. Most of the interference structures, including all those interpreted as sheath folds, are wholly D_1 in origin and are not the result of superposition of F_2 folds on F_1 folds. Removal of the D_2 strain (Schultz-Ela and Hudleston, 1991) does not remove the sheath folds. (3) Basalt-cored anticlines southeast of the Leech Lake stratotectonic disconformity (fig. 3; also, Jirsa and others, 1992) are regional F_1 structures in which sequences of relatively massive metavolcanic rocks are overlain by massive to thin-bedded sedimentary strata. The accommodation of the massive rocks in the cores to folding under conditions that did not produce a fabric in overlying sedimentary strata must have involved brittle slip on numerous fractures, and further implies that many, if not most, of the contacts between the volcanic and sedimentary sequences in these anticlines were decollements. Field study affirms that many of these contacts are indeed slip surfaces (Southwick, 1991a), but the time at which shearing occurred (D_1, D_2, or both) has not been resolved.

The current working hypothesis for F_1 folding in the Minnesota part of the Wawa subprovince is that it was a consequence of subduction. The folds are thought to have formed at a moderate depth in an accretionary wedge where sedimentary materials were unlithified, probably wet, and capable of deforming by boundary-slip processes (MacKay and Moore, 1990; DiTullio and Byrne, 1990). More massive and drier volcanic rocks in the same environment would have deformed also by internal slip on pillow margins, fragment boundaries, and other primary and secondary fracture surfaces. Slip on the various surfaces may have been eased by the incipient growth of secondary hydrous minerals (clay minerals and chlorite, particularly) along them.

The structures observed in the isolated Ishpeming greenstone belt in Upper Michigan are comparable to those in the Minnesota segment of the Wawa subprovince, but they differ in detail (Johnson and Bornhorst, 1991). In Michigan, the first folds (F_1) are recumbent structures that have a strong axial-planar fabric and involve rocks that locally record amphibolite-facies metamorphism. Therefore, F_1 folds evidently formed at deeper levels in Michigan then they did in Minnesota, if in fact they are strictly correlative structures. The F_2 folds in Michigan, as in Minnesota, are upright and somewhat oblique to F_1. A strong tendency exists for F_2 folds to exhibit Z-asymmetry, which suggests that a component of dextral shear may have been involved in their formation.

The principal new finding with regard to the stratigraphic evolution of Archean volcanic sequences in the Lake Superior region is the recognition of a fundamental stratotectonic disconformity in the midst of a succession formerly viewed as conformable, in the Wawa subprovince of northern Minnesota (Jirsa and Boerboom, 1990; Jirsa and others, 1992). This break (LLSD, fig. 4) is apparently a stratigraphic hiatus that later became a favored zone for faulting. It separates volcanic sequences of contrasting composition. Rocks south of the break comprise a geochemically evolved mafic sequence and a thick dacitic sequence that is in part pyroclastic and in part epiclastic. Mafic rocks north of it are as a whole more primitive geochemically, and the proportion of dacitic material in the volcanic sequence is substantially smaller. Structural contrasts across the break also are evident. The large, recumbent F_1 folds just discussed are prominent south of it, whereas the rocks on the north side appear to have been imbricated by faulting prior to (and during) D_2, without extensive development of early fold nappes.

Perhaps the most significant aspect of this newly recognized disconformity is its large regional extent. It can be traced geophysically for a strike length of more than 430 km in Minnesota (figs. 3, 4), and therefore it is a major element that must be taken into account in tectonic syntheses and economic evaluation of the subprovince.

The "quiet zone" at the south margin of the Wawa belt has received scant attention since its original recognition in aeromagnetic data (Chandler and Southwick, 1990). The few samples retrieved from beneath thick glacial cover in this area (Southwick and others, 1986; 1990) are rock types that are well known in the Wawa subprovince. However, they have undergone a feeble alteration process that has produced much secondary epidote, albite, and quartz, and has consumed oxides and primary mafic silicates. The extent, cause, and significance of this regional propylitization remain intriguing questions.

CONCLUSIONS

1. The subprovince concept as devised by Card and Ciesielski (1986) and Card (1990) for subdividing the Superior province of the Canadian Shield is a useful frame of reference in which to describe and interpret the Archean rocks of the United States segment of the Lake Superior region. The concept is especially applicable to the Archean terranes of Minnesota.

2. Structural observations indicate that the main Late Archean deformational events throughout the southern Superior province developed in a regime of dextral transpression. Deformation was diachronous in Canada, and in a general way progressed from north to south as a function of time (Card, 1990, and references therein). This pattern of southward tectonic younging has not yet been confirmed by detailed geochronology in the United States.

3. The Quetico-Wabigoon subprovince boundary in Minnesota is a zone of large-scale dextral faulting across which neither rock sequences nor metamorphic grade correlate. Although structural sequences and styles are similar on either side of the boundary, geochronologic data indicate that deformation was not synchronous. Thus, the Quetico-Wabigoon boundary can be interpreted as a boundary between stratotectonic terranes.

4. The Quetico-Wawa subprovince boundary is not everywhere a fault contact, although major faults do lie on it or near it in many places. Structural sequences and strain patterns are consistent with synchronous dextral transpression across the boundary zone. Differences in the development of D_1 fabric on either side of the boundary are reasonably attributed to differences in the depth and metamorphic environment during F_1 folding. Compelling evidence is lacking that the Quetico-Wawa join represents a boundary between stratotectonic terranes.

5. Although it probably was deformed by dextral transpression together with its surroundings, the interior of the Quetico subprovince records little evidence of shear strain. Failure to record shearing is attributed to the mechanical consequences of episodic but extensive emplacement of granitic material into the core of the Quetico subprovince through much of its tectonic evolution.

6. Early folding (F_1) in the Minnesota segment of the Wawa subprovince was recumbent in style and occurred at shallow levels under conditions where deformation took place mainly by grain boundary slip. Later upright folding (F_2) produced strong deformation fabrics under deeper seated metamorphic conditions. The larger F_1 folds are of kilometer and larger size and determine the distribution of rock units in the Vermilion district and its southwestward extensions. F_1 folding preserved in the Ishpeming belt in Michigan apparently took place at greater depths than the F_1 folding in northeastern Minnesota.

7. A major stratotectonic disconformity in the Wawa subprovince of Minnesota separates two belts of contrasting volcanic stratigraphy and early-stage structural evolution. Large F_1 folds are developed south of the disconformity, in mafic rocks of evolved petrochemistry and in associated sedimentary sequences of dacitic provenance. The belt north of the disconformity lacks large F_1 structures and includes volcanic units that are, as a whole, less evolved geochemically.

8. The Vermilion fault in northern Minnesota is a relatively young Archean structure that is oblique to lithologic contacts along much of its trace, and offsets the traces of other faults (Haley, Rainy Lake–Seine River) that have ductile or ductile-brittle attributes. It is not a terrane-bounding structure.

9. The "quiet zone" at the south edge of the Wawa belt is a zone in which weak hydrothermal alteration has occurred on a regional scale. The alteration event is neither well characterized nor clearly understood. The Great Lakes tectonic zone at the south margin of the quiet zone is the complex bounding surface between terranes of distinctly different rock type, structural style, and tectonic history as recorded geochronologically. Therefore, the Great Lakes tectonic zone meets the criteria for a stratotectonic terrane boundary.

MINNESOTA RIVER VALLEY SUBPROVINCE (ARCHEAN GNEISS TERRANE)

By P.K. Sims

Rocks assigned to the Minnesota River Valley subprovince, previously called the Archean gneiss terrane, form a substantial part of the basement of southwestern Minnesota (Chandler and Southwick, 1990; pl. 1), south of the Great Lakes tectonic zone. They are exposed in the Minnesota River Valley, where erosion has cut below a thin cover of Cretaceous sedimentary rocks (Grant, 1972; Bauer and Himmelberg, 1993). Scattered exposures of these Archean gneissic rocks occur elsewhere in east-central Minnesota, northern Wisconsin, and the Upper Peninsula of Michigan, mainly in the cores of gneiss domes or uplifted fault blocks that owe their exposure to Early Proterozoic Penokean deformation. Presumably these gneisses formed a continuous crust (protocontinent) of wide areal extent prior to Early Proterozoic continental rifting and the younger Middle Proterozoic rifting that formed the Midcontinent rift system. Archean gneisses exposed in central Wisconsin (Sims and others, 1989), in the Early Proterozoic Marshfield arc terrane (of Wisconsin magmatic terranes), are not contiguous with the larger crustal segment. This segment (mini-continent) was rafted in with arc crust during continent-arc convergence in the Early Proterozoic Penokean orogeny.

SOUTHWESTERN MINNESOTA

The gneissic rocks of the Minnesota River Valley subprovince are best represented by outcrops in its largest and most coherent part, the Minnesota River Valley; they consist of complex migmatitic felsic gneisses, local schistose to gneissic amphibolite, and lesser metagabbro and metasedimentary gneiss. Most of the rocks are older than 3,000 Ma and have undergone a protracted history of multiple deformation and metamorphism that culminated prior to emplacement of the 2,600 Ma late- to post-tectonic Sacred Heart Granite.

Recent geophysical studies of these southwestern Minnesota Archean gneisses have shown that they constitute four major blocks (pl. 1) of different age, rock types, and metamorphic grade, and that geophysical boundaries between blocks are probably faults that dip northward (Chandler and Schaap, 1992). Possibly these blocks were accreted separately at different times.

In all four blocks (pl. 1), felsic gneisses are the most abundant rock type. In the Morton block, the gneiss is a tonalitic to granodioritic migmatite derived from igneous protoliths and intruded by several types of granitic gneisses. In the Montevideo block, three types of felsic gneiss exist: (1) banded granodiorite gneiss, (2) coarse-grained granite gneiss, and (3) medium-grained granite gneiss. Biotite-garnet gneiss of graywacke derivation in the Montevideo block also contains orthopyroxene, quartz, plagioclase, and minor potassium feldspar and cordierite (Grant, 1972; Moecher, 1984).

The principal Archean granitoid rock, the Sacred Heart Granite (Lund, 1956), cuts the gneisses and is virtually undeformed. It is typically pink, medium grained, and homogeneous to weakly foliated. It has a Pb-Pb age of 2,605±0.006 Ma, and its lead isotope characteristics are typical of mesozonal late-tectonic plutons (Doe and Delevaux, 1980). Although sparsely exposed in the Minnesota River Valley, granitoid rocks are moderately abundant in the covered basement, as determined from drilling (Sims, 1990d). Associated gneisses penetrated by drilling in covered areas are lithologically similar to those exposed in the river valley (Southwick and others, 1988).

The gneisses in the Minnesota River Valley are polydeformed (Bauer and Himmelberg, 1993). The dominant foliation (S_1) is folded by small-scale folds (F_1), but the dominant folds—large-scale, upright antiforms and synforms that plunge gently east—are F_2 structures. Bauer (1980) has recognized minor F_3 and F_4 folds in the Montevideo block. Small, conjugate ductile shear zones in the Montevideo block postdate the F_3 folds.

The gneisses throughout the Minnesota River Valley were generally metamorphosed under the same high-grade, granulite-facies conditions (Himmelberg and Phinney, 1967; Grant, 1972). Geothermobarometry from throughout the valley yields temperatures of 650–750 °C and pressures of 4.5–7.5 kbar (Leier and Perkins, 1982; Perkins and Chippera, 1985; Moecher and others, 1986). Moecher and others considered 700±50 °C and 6.0±1.5 kbar to be the best estimate for peak granulite-facies metamorphic conditions; they attributed the variations to reequilibration of assemblages with falling temperatures and to differences among calibrations applied. From those data, Moecher and others calculated a Late Archean geotherm of 32 °C/km for the Minnesota River Valley.

Retrograde mineral assemblages of amphibolite and greenschist facies are locally common in the gneisses. The retrogression possibly resulted from discrete lower grade metamorphic events.

Extensive geochronologic investigations by S.S. Goldich and colleagues have defined four major events, although Rb-Sr and U-Pb systems have been disturbed (Goldich, Hedge, and others, 1980; Goldich and Wooden, 1980). The intrusive protolith of the tonalite-granodiorite gneisses is ≈3,500 Ma; high-grade metamorphism of these gneisses at ≈3,050 Ma was followed by high-grade metamorphism at ≈2,600 Ma, followed by emplacement of the Sacred Heart

Granite. In the Early Proterozoic, a low-grade thermal event at ≈1,840 Ma (Doe and Delevaux, 1980; Goldich and others, 1970) was accompanied by the emplacement of small mafic to felsic igneous bodies.

An isotopic age (2,624±57 Ma) of tonalitic basement gneiss in the eastern part of the Jeffers block (pl. 1), together with a Sm-Nd model age (≈2,700 Ma) on the same sample, suggests that the gneisses in this southerly block are Late Archean and formed from a crustal source that separated from the mantle in the Late Archean (Southwick and others, 1994). Because this gneiss (gneissic tonalite) shows no evidence that it interacted with 3,600 Ma continental crust, Southwick and others (1994) have suggested that the Morton block may comprise the southernmost segment of Early Archean crust and the Jeffers block may comprise Late Archean crust.

Differential uplift of the Archean gneisses of southwestern Minnesota during the Early Proterozoic is indicated by K-Ar and Rb-Sr biotite ages (Goldich and others, 1970). Biotite from gneisses in the Montevideo and Benson blocks (pl. 1) has ages of about 1,850 Ma, which is approximately the same age as a small granite pluton in the same area (Section 28 granite of Goldich and others, 1970). Biotite ages from gneisses in the Morton block, however, range from 2,580 to 2,300 Ma, indicating that this block stabilized before 2,300 Ma. Uplift of the Montevideo and Benson blocks in Early Proterozoic time presumably took place along the fault that separates the Montevideo and Morton blocks (pl. 1); it probably was caused by compressive tectonism during the Penokean orogeny.

PENOKEAN GNEISS DOMES AND FAULT BLOCKS

The Archean gneiss exposed in the cores of Early Proterozoic (Penokean) gneiss domes and uplifted fault blocks is more diverse lithologically and temporally than the gneisses in southwestern Minnesota and includes supracrustal rocks of definite Late Archean age. Except for the Watersmeet dome (loc. 4, fig. 5), rocks older than 3,000 Ma have not been found. Another difference is the extensive reworking of the gneisses during the Penokean deformation that produced the gneiss domes at Watersmeet, Mich. (4, fig. 5), and McGrath, Minn. (1, fig. 5). In contrast, the fault-bounded uplifts in northern Michigan, such as the southern complex of the Marquette district (loc. 6, fig. 5) were virtually undeformed during the Penokean orogeny. Cannon (1973) first pointed out that Early Proterozoic metagabbro dikes that cut gneisses in the southern complex are undeformed, and subsequent strain analysis studies (Cambray, 1984) have confirmed this observation. A summary of the lithology, structural characteristics, and age of Archean gneisses in the domes and fault uplifts is given in table 1.

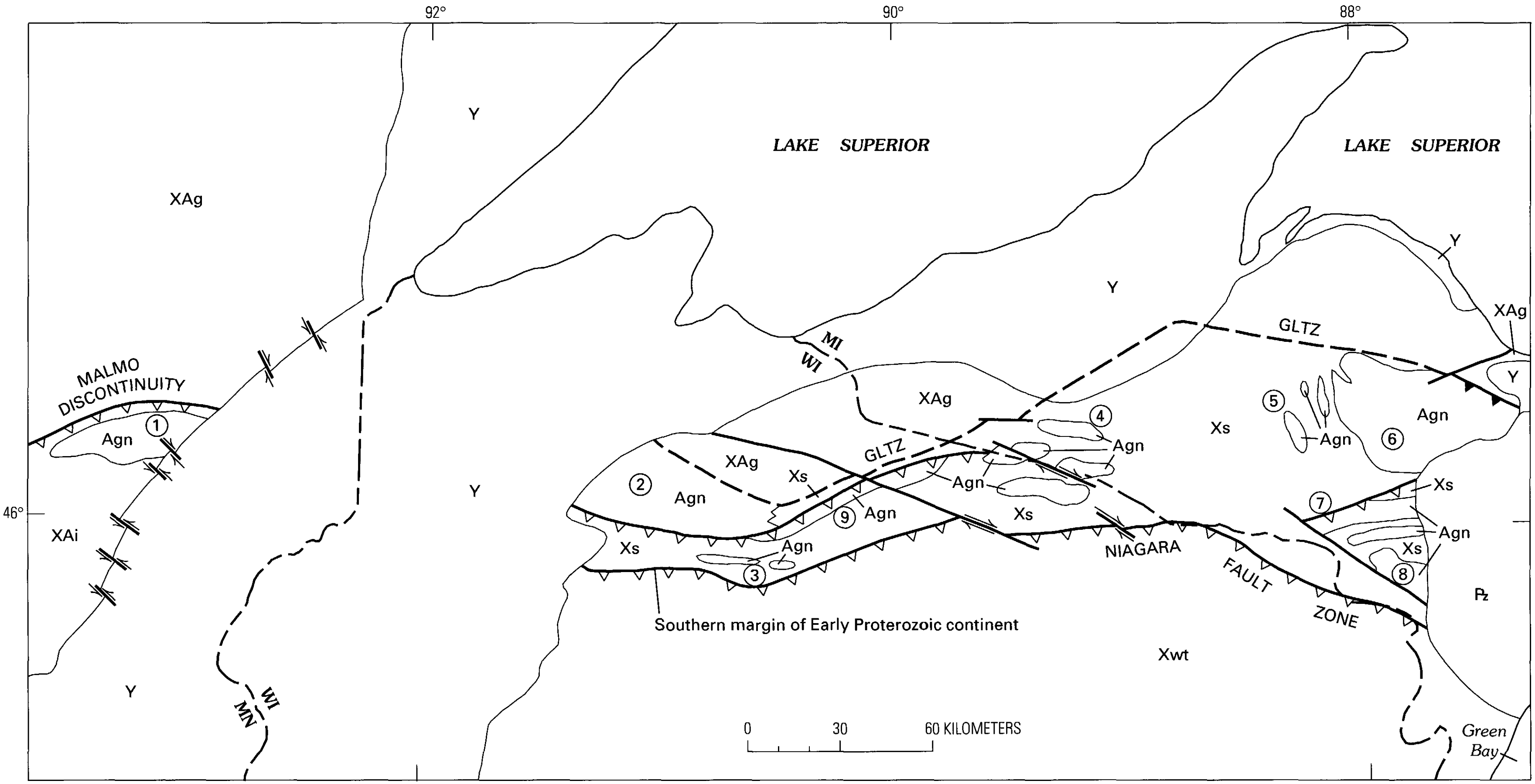
EAST-CENTRAL MINNESOTA

East-central Minnesota contains poorly exposed gneisses, some of which at least are Archean in age, and several types and ages of granitoid rocks of Early Proterozoic age. Except for the McGrath Gneiss of definite Archean age (loc. 1, fig. 5), the metamorphic and granitoid rocks are included in unit XAi in figure 5 and on plate 1.

The McGrath Gneiss comprises a well-documented gneiss dome flanked by metamorphosed supracrustal rocks assigned to the Early Proterozoic Denham Formation (Morey, 1978; Holm, 1986). It dominantly comprises a pinkish-gray medium- to coarse-grained, locally migmatitic quartzofeldspathic augen gneiss. Both ductile and brittle structures are widespread in the gneiss, and ductile deformation is particularly well developed along the contact between the gneiss and the overlying supracrustal Early Proterozoic rocks. The McGrath Gneiss has a Rb-Sr whole-rock isochron age of ≈2,700 Ma (Stuckless and Goldich, 1972; Keighin and others, 1972). Biotite ages (both Rb-Sr and K-Ar methods) from cataclastic zones in the gneiss are similar to a Rb-Sr whole-rock age of 1,740–1,730 Ma for Early Proterozoic rocks in the area (Keighin and others, 1972), indicating that the doming and attendant ductile and brittle deformation took place in Early Proterozoic time.

The Hillman Migmatite, included in unit XAi (fig. 5), also is interpreted as a dome (or domes), particularly in the immediate vicinity of relatively large Early Proterozoic intrusions (Southwick and others, 1988). The paleosome of the migmatite consists dominantly of dark-gray, fine- to medium-grained biotite-garnet-cordierite schist, hornblende schist, and metagraywacke, probably mainly of sedimentary origin (Dacre and others, 1984). The neosome is a gray, medium- to coarse-grained, vaguely foliated equigranular tonalite that is conformably interlayered on various scales with the paleosome. The age of the Hillman is uncertain; it could be Archean, Early Proterozoic, or a mixture of the two (Morey, 1978). Goldich and Fischer (1986) obtained a U-Pb zircon age of 1,869±5 Ma on a relatively massive foliated tonalite that intrudes the Hillman Migmatite, but this tonalite is not necessarily the same age as the tonalitic neosome in the migmatite, as inferred earlier by Goldich and Fischer (1986).

Other gneissic units in east-central Minnesota of possible Archean age are the Richmond Gneiss of Morey (1978) and the Sauk Rapids Metamorphic Complex of Morey (1978). The Sartell Gneiss of Morey (1978) (of the Sauk Rapids Metamorphic Complex) contains mineral assemblages (orthopyroxene, garnet, and cordierite) diagnostic of the granulite facies. Mineral-pair thermobarometry suggests an approximate temperature and pressure of metamorphism of 800 °C and 5.1 kbar (Dacre and others, 1984). Although isotopic ages are lacking for the gneisses, the granulite-facies metamorphism coupled with the presence of relatively low pressure cordierite in the rocks favors an Archean age for these gneisses. Alternatively, the gneisses could be



EXPLANATION

Pz	Rocks of Paleozoic age
Y	Rocks of Midcontinent rift system (≈1,100 Ma)
Xs	Metasedimentary and metavolcanic rocks of continental margin assemblage (Marquette Range Supergroup)
Xwt	Wisconsin magmatic terranes (1,835–1,890 Ma)
XAi	Metamorphic and granitoid rocks of Early Proterozoic and Archean age
XAg	Early Proterozoic supracrustal rocks and rocks of Archean greenstone-granite (Wawa subprovince) terrane
Agn	Archean gneiss of gneiss terrane (Minnesota River Valley subprovince)

Contact

Transcurrent fault—Barbs show relative horizontal movement

Thrust fault of Early Proterozoic age—Sawteeth on upper plate

Thrust fault of Late Archean age—Sawteeth on upper plate

GLTZ **Great Lakes tectonic zone** —Dashed where covered by Proterozoic rocks

① **Locality referred to in text**

Figure 5. Geologic map showing Archean-cored gneiss domes and uplifts, Lake Superior region. Compiled by P.K. Sims, 1991.

Table 1. Lithologic and structural characteristics and age of Archean gneisses in domal structures and fault-bounded uplifts, Minnesota River Valley subprovince.

Locality and designation in figure 5	Formation	Rock types	Internal structure	External structure	Rock texture	Age, in Ma	References: Geology	References: Geochronology
McGrath dome, 1.	McGrath Gneiss	Granitic gneiss	Generally flat foliation	Gneiss dome	Granoblastic; ductile and brittle shear zones.	≈2,700	Morey, 1978	Stuckless and Goldich, 1972.
Gneisses near Morse, 2.	Unnamed	Granite gneiss and amphibolite.	Foliation, WNW., steep Lineation, gentle	No data	Protomylonite	No data	Sims, Peterman, and others, 1985.	
Fifield, 3	Unnamed	Felsic biotite gneiss	Foliation, E.-W., vertical.	Unknown	Granoblastic	2,984±51		W.R. Van Schmus, written commun., 1988.
Watersmeet dome, 4.	Unnamed	Leucogranite Interlayered amphibolite and biotite gneiss Layered felsic gneiss Tonalitic augen gneiss	Archean foliation, E.-W., steep. Proterozoic foliation, NE., moderate.	Gneiss dome	Granoblastic, cataclastic.	2,585 2,640-2,750 3,562 3,562	Sims and others, 1984.	Sims and others, 1984; Peterman and others, 1980; Peterman and others, 1986.
Amasa uplift, 5.	Margeson Creek Gneiss.	Layered tonalitic gneiss Megacrystic granite gneiss Amphibolite	Foliation, NW., steep Lineation, gentle	Gneiss dome or fault-bounded uplift.	Protomylonite	Late Archean	Gair and Wier, 1956; Foose, 1981.	
Southern complex, 6.	Compeau Creek Gneiss. Bell Creek Gneiss.	Migmatitic felsic gneiss. Megacrystic granite gneiss, Amphibolite, iron-formation.	Foliation, dominantly NW., steep. Lineation, moderate to steep.	Gneiss dome or fault-bounded uplift.	Granoblastic, weak cataclastic.	≈2,775	Cannon and Simmons, 1973; Taylor, 1967.	W.R. Van Schmus, oral commun., 1982; Hammond, 1978.
Dickinson County, 7.	Unnamed	Gneiss and granite				≈2,800-2,600	James and others, 1961.	W.R. Van Schmus, oral commun., 1982.
	Dickinson Group	Conglomerate, schist and amphibolite.	Foliation, E.-W., steep Lineation, ≈25° S. 80° E.	Gneiss dome(?) modified by thrust.	Granoblastic, protomylonite.	>2,600		
	Granite gneiss of the Norway Lake area.	Augen gneiss			Protomylonite cataclasite.	≈2,600		
Carney Lake dome, 8.	Carney Lake Gneiss.	Gray tonalitic gneiss Red granitic gneiss	Foliation, NW., steep Lineation, gentle E.	Gneiss dome	Protomylonite	≈2,750	Bayley and others, 1966.	Van Schmus and others, 1978.
Flambeau River, 9.	Unnamed	Interlayered biotite gneiss and amphibolite.	Foliation, NE., steep Lineation, steep	Thrust block	Granoblastic, protomylonite, cataclasite.	≈2,800		W.R. Van Schmus, oral commun., 1982.

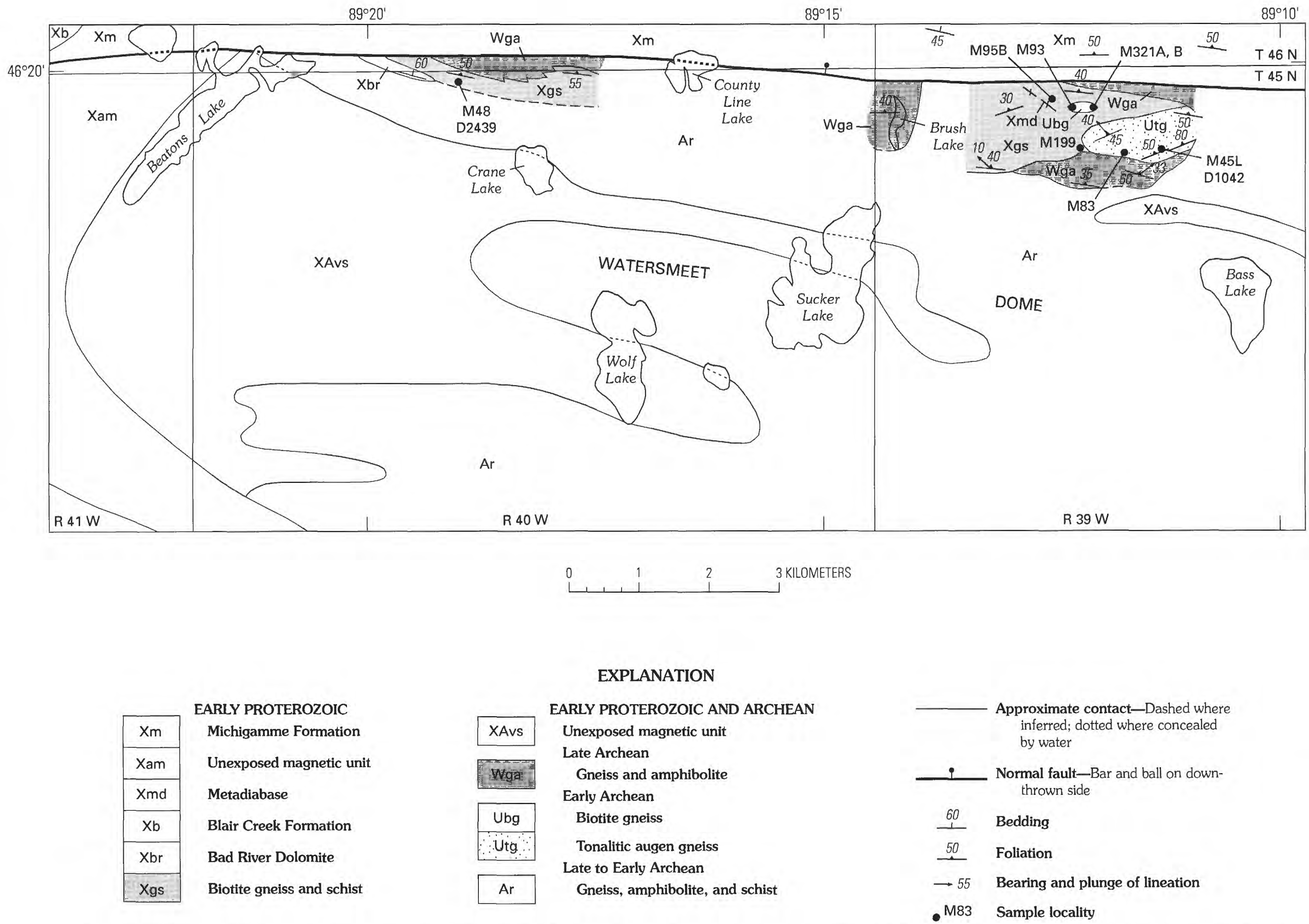


Figure 6. Geology of northern part of Watersmeet gneiss dome, northern Michigan. No contact line between units indicates position of contact indeterminate. Modified from Sims (1990e).

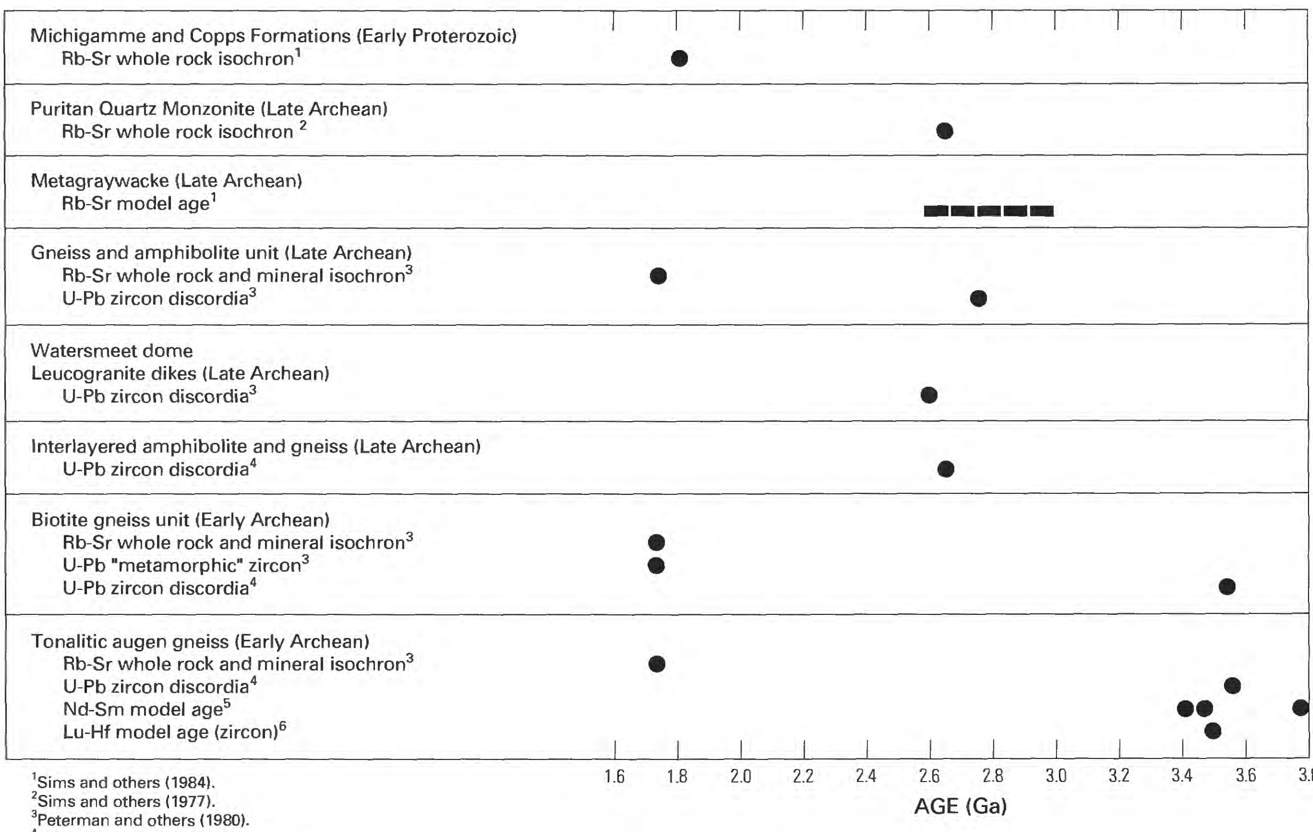
approximately contemporaneous with those in the Pembine-Wausau terrane in northeastern Wisconsin, such as the Dunbar Gneiss (age, 1,862±5 Ma; Sims, Peterman, and others, 1985) and the Marinette Quartz Diorite (age, 1,857±6 Ma; Sims and Schulz, 1993). The gneisses in Wisconsin are intruded by younger foliated to unfoliated granitoid rocks, in the same manner as those in east-central Minnesota.

WATERSMEET GNEISS DOME

The Watersmeet gneiss dome (loc. 4, fig. 5) is an elliptical structure 24 km long (east-west) and 7.5 km wide that has been delineated mainly by aeromagnetic and gravity data (Klasner and Sims, 1984). Archean gneisses and schists in the core are flanked by Early Proterozoic rocks assigned to the Marquette Range Supergroup (fig. 6). An unexposed magnetic unit believed to be correlative with the Blair Creek Formation overlies the gneisses on the south and west; elsewhere, the Michigamme Formation directly overlies the gneisses. The north margin is a high-angle fault (Sims, 1990e). Exposures in the dome are sparse and are confined to three small areas along the north margin. The dome is a short distance southeast of the Great Lakes tectonic zone (fig. 5). A rim syncline in the Michigamme Formation 1–2 km outside the core has been delineated on the north and east sides of the dome (Sims, 1990e).

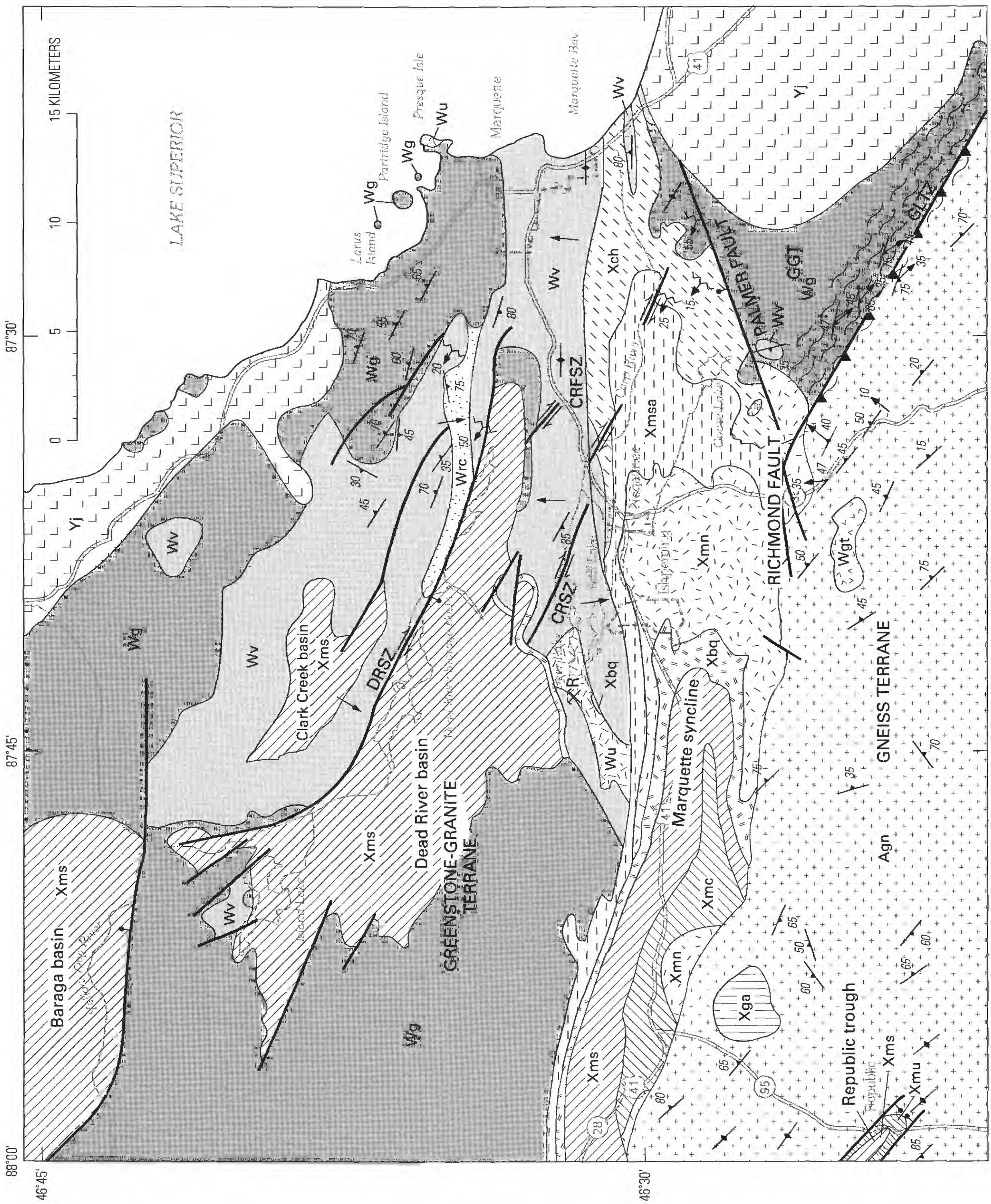
The Watersmeet dome formed at a higher pressure (7 kbar) than the other gneiss domes in Michigan (fig. 5) and at temperatures in the range of 600–650 °C (Attoh and Klasner, 1989). As a consequence, rocks in the core of the dome were metamorphosed to amphibolite facies and were complexly folded and tectonically interleaved and mylonitized during the doming. Secondary Rb-Sr whole-rock and mineral isochrons on the Archean rocks give ages of 1,800 to 1,750 Ma, which presumably represent the time of metamorphic recrystallization during the doming event (Sims and others, 1984; Peterman and others, 1986). The doming occurred more than 50 m.y. after the culmination of collision along the continent-arc boundary (Niagara fault zone), which has been dated at approximately 1,850 Ma (Sims and others, 1989).

Three major Archean units have been delineated in the Watersmeet dome (fig. 6). From oldest to youngest, these are (1) tonalitic augen gneiss (unit Utg); (2) a complexly interlayered biotite gneiss succession (unit Ubg), which overlies (tectonically?) the tonalitic augen gneiss; and (3) an interlayered bimodal amphibolite and biotite gneiss succession (unit



[1]Sims and others (1984).
[2]Sims and others (1977).
[3]Peterman and others (1980).
[4]Peterman and others (1986).
[5]McCulloch and Wasserberg (1980); Futa (1981).
[6]Patchett and others (1981).

Figure 7. Summary chart of radiometric ages in the Watersmeet area, northern Michigan. Dashed bar indicates the uncertainty in the model age. Modified from Peterman and others, 1986.



LAKE SUPERIOR
0 5 10 15 KILOMETERS
87°30'
87°45'
88°00'
46°45'
46°30'
Larus Island
Partridge Island
Presque Isle
Marquette
Marquette Bay
PALMER FAULT
RICHMOND FAULT
GGT
GLTZ
CRFSZ
CRSZ
DRSZ
Clark Creek basin
Dead River basin
Baraga basin
GREENSTONE-GRANITE TERRANE
Marquette syncline
GNEISS TERRANE
Republic trough
Republic
Yj
Wg
Wv
Wu
Wrc
Wgt
Xms
Xmsa
Xch
Xmn
Xbq
Xmc
Xga
Xmu
Agn
41
28
95

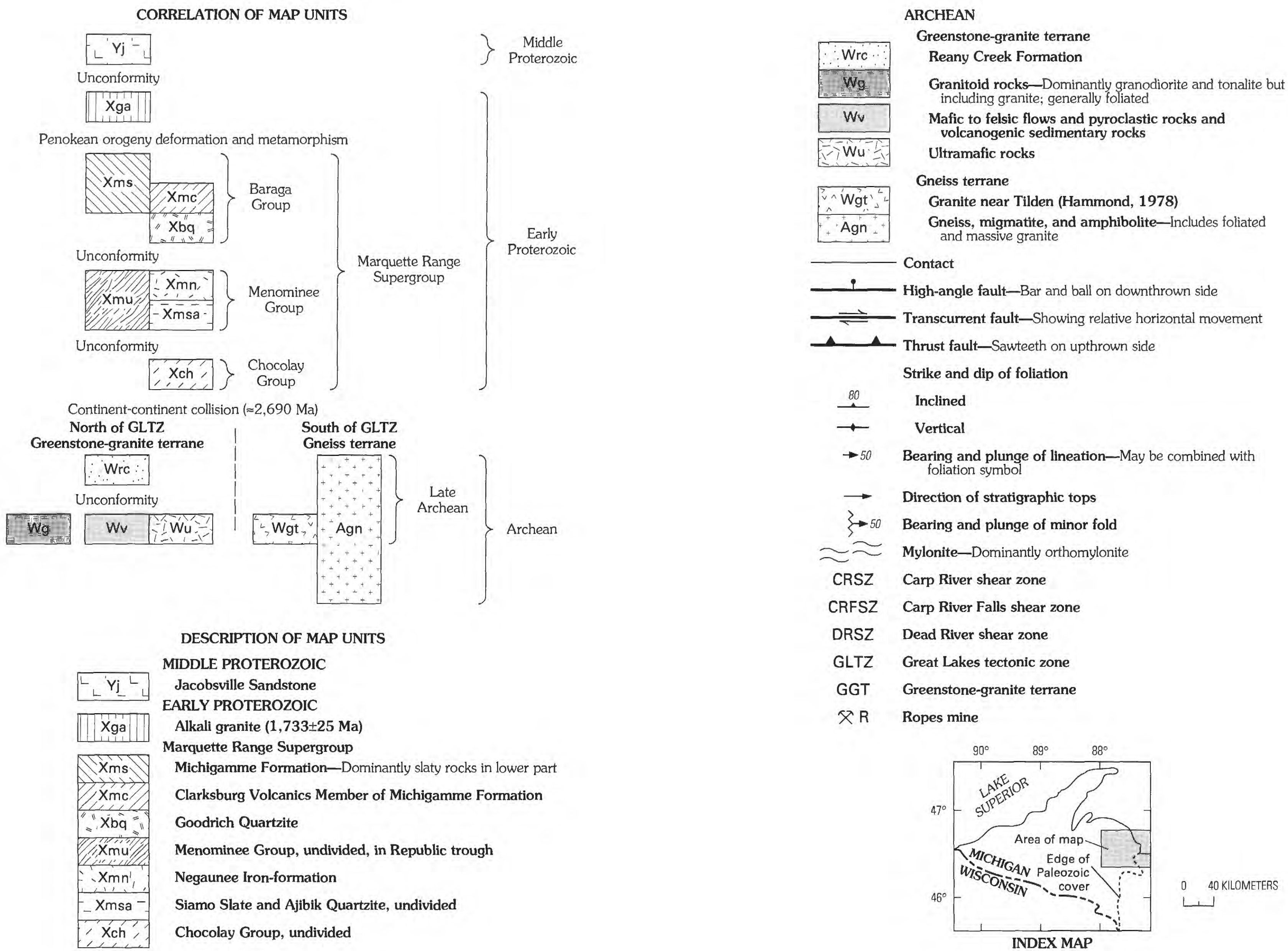


Figure 8. Geologic map of part of the Marquette 1°×2° quadrangle, showing Great Lakes tectonic zone. From Sims (1991a).

Wga) (Sims, 1990e). In addition, Early Proterozoic rocks within the dome include a fine-grained biotite schist (unit Xgs) and, at least locally, the Bad River Dolomite (unit Xbr). The fine-grained schist resembles garnetiferous schist of the Michigamme Formation. It differs petrographically from the Archean biotite gneiss and schist in having as much as 6 percent muscovite and as much as 1 percent fluorite. During mapping (Sims and others, 1984), this rock was considered to be Archean, but subsequent Nd isotopic studies showed it to be Early Proterozoic (Barovich and others, 1991), as discussed following. Because of its moderately high radioactivity, the rock unit has been investigated by private mining companies as a possible source of U, Th, and rare earth elements.

The Nd data (Barovich and others, 1991) indicate that Archean units in the Watersmeet dome have Early Archean T_{CHUR} ages between 3.72 and 3.54 Ga. These model ages are in agreement with the U-Pb upper intercept age of 3.56 Ga determined for the tonalitic augen gneiss and biotite gneiss (Peterman and others, 1980). At the U-Pb crystallization age of 3.56 Ga, ε_{Nd} values for these samples range from –2.0 to +0.2. The Early Proterozoic biotite schist unit (Xgs), on the other hand, has T_{DM} ages between 2.37 and 2.25 Ga. Nd and Sm concentrations in this unit are as much as 4.5 times higher than average upper crustal felsic material (Barovich and others, 1991).

A detailed isotopic age study of the rocks in the Watersmeet dome and adjacent areas indicates three major rock-forming and tectonic events (Peterman and others, 1986; fig. 7): (1) 1.82–1.72 Ga, the uplift age of the dome, (2) 2.75–2.60 Ga, and (3) 3.60–3.50 Ga. The tonalitic augen gneiss and the biotite gneiss unit formed in the interval 3.60–3.50 Ga. Evidence is lacking for thermotectonic events in the interval from 3.6 Ga to the Late Archean. A Late Archean event, 2.75–2.60 Ga, comprised (1) bimodal volcanism, (2) clastic sedimentation, (3) folding and metamorphism, and (4) granitic plutonism. The bimodal volcanic rocks were deposited on an older gneiss basement. Folding on steep axes occurred before the emplacement of small bodies of leucogranite (2.59 Ga). The geologic record between the time of stabilization of the Archean crust (≈2.5 Ga) and deposition of the Marquette Range Supergroup is lacking in the area, but an upper age of 1.9 Ga or older for its deposition commonly is accepted, based largely on ages of volcanic rocks in Wisconsin (W.R. Van Schmus, written commun., 1980). In the Watersmeet dome, these supracrustal rocks were deformed and metamorphosed between 1.78 and 1.72 Ga, as a consequence of ductile deformation concurrent with uplift. Rb-Sr whole-rock and mineral ages were reset by this event.

SOUTHERN COMPLEX OF MARQUETTE DISTRICT

The southern complex, in the Upper Peninsula of Michigan, is a large, northwest-oriented domal structure that closes to the northwest and was uplifted during the Penokean deformation. It is more than 55 km long and a maximum of 32 km wide (loc. 6, fig. 5). It is covered to the east by Paleozoic sedimentary rocks. Exposures are moderately abundant in the northern part of the body but are virtually absent in the southern part. The southern complex mainly comprises gneiss and amphibolite of the Minnesota River Valley subprovince (Archean gneiss terrane), but that part north of the Great Lakes tectonic zone, in the area south of Marquette (fig. 8), consists of volcanic and granitoid rocks of the Ishpeming greenstone belt that belong to the Wawa subprovince (Superior province). The origin of the domal structure is not well understood. The northwest nose of the structure is mantled by Proterozoic rocks and is indented by an infolded belt (Republic trough) of these rocks; the Proterozoic rocks dip steeply and at least in part are interpreted as being in fault contact with the Archean gneisses (Cannon, 1975). Regardless of the origin of the domal structure, the Archean gneisses in its core were neither appreciably deformed nor metamorphosed during the Early Proterozoic deformation. Structures formed during Archean polydeformation of the gneisses are well preserved except immediately adjacent to the Marquette syncline (fig. 8), where they are overprinted by brittle fracturing and locally by a crenulation cleavage.

The gneisses in the northwestern part of the domal structure consist of foliated granitoid rocks and generally older migmatite and banded gneiss (Cannon, 1986). The foliated granitoids are mostly granodiorite, but locally are tonalite and granite (Cannon and Simmons, 1973) that have a prominent secondary foliation (S_2). In one rock type (Bell Creek Gneiss), the foliation is expressed by aligned microcline megacrysts, but in other rock types it is mainly expressed by shears along which micaceous minerals are aligned. In places a subtle compositional layering is parallel to the foliation. The migmatite and banded gneisses have a prominent compositional layering of (1) coarse- to medium-grained tonalite and granodiorite, (2) amphibolite, and (3) dark-colored biotite-rich quartzofeldspathic rocks. In the migmatite, the leucosome is granitic in composition and partly discordant to the foliation. At several localities, lenses of highly metamorphosed iron-formation containing grunerite, iron-rich orthopyroxene, clinopyroxene, garnet, magnetite, and quartz occur in mafic rocks (Cannon and Simmons, 1973).

In the northwest nose of the antiformal structure, Taylor (1967) determined two principal phases of Archean deformation: (1) early, probably flat-lying folds with axial planes (S_1) trending northeastward, and (2) younger upright folds having steep northwest-trending axial surfaces. F_1 structures are observed only in the migmatite and banded gneisses. F_2 mainly controls the distribution of the rock units, and the axial-plane foliation (S_2) is the most prevalent structure in the rocks.

In the area south of Marquette (fig. 8), compositionally layered medium-grained gneiss and migmatite are the dominant rock types in the gneiss terrane (Sims, 1991a). The layered gneisses range in composition from tonalite to granite and contain variable amounts of interlayered medium-grained amphibolite. The layering and compositions indicate a probable metavolcanic origin for those rocks. Pink aplitic granite and granite pegmatite commonly transect the felsic gneiss and amphibolite and locally form migmatite. A pinkish-gray, medium-grained, massive to weakly foliated granite, informally called the "Tilden granite" (Hammond, 1978), intrudes the gneisses at places. It is similar in composition and texture to the Sacred Heart Granite.

In the Marquette area, early gently inclined to recumbent folds oriented northwestward plunge gently to moderately northward and are the dominant structure in the gneisses (Sims, 1991a). A younger steep, northwest-oriented foliation (Sims, 1993) such as that present in the Republic area has been recognized only locally in the area, although reversals in the dip of the foliation suggest the existence of broad, open, younger northwest-oriented folds. The folding and metamorphism preceded emplacement of the Tilden granite of Hammond (1978).

A sample of gray tonalitic gneiss from the Marquette area has a U-Pb zircon age of 2,779±21 Ma and a lower intercept age of 802±76 Ma (Hammond, 1978; recalculated by Zell E. Peterman). A sample of the Tilden granite of Hammond (1978) has a U-Pb zircon age of 2,585±15 Ma (R.E. Zartman, written commun., 1991).

TECTONIC ENVIRONMENT

The gneiss domes and uplifts occur within a 70-km-wide 500-km-long east-west belt north of the Niagara fault zone (pl. 1; fig. 5), and resulted from collision of the rifted passive margin and the Early Proterozoic Pembine-Wausau terrane of the Wisconsin magmatic terranes. The belt comprises most of the fold-and-thrust belt of the Penokean orogeny, and is characterized by mainly south dipping, north-verging thrust faults (Attoh and Klasner, 1989; Klasner and Sims, 1993; Gregg, 1993). As discussed by Attoh and Klasner (1989), tectonic stacking on Archean continental crust could have caused the heating required for melting and remobilization of gneissic basement, as especially well manifested by the Watersmeet dome. Pressure-temperature indicators in the Watersmeet dome require a depth in excess of 25 km for the metamorphism.

Gregg (1993) has compared the foreland fold-and-thrust belt in northern Michigan with that in the Northern Appalachians, pointing out that the progression from parautochthonous rocks at the northern tectonic front, to the gneiss domes (metamorphic core complexes), and finally to the Niagara suture shows an arrangement remarkably similar to that of the Northern Appalachians.

GREAT LAKES TECTONIC ZONE

By P.K. Sims

The Great Lakes tectonic zone (GLTZ) is an Archean crustal boundary of subcontinental length that separates the Minnesota River Valley subprovince from the Wawa subprovince (Superior province) on the north (Sims and others, 1980). Thus, it separates the gneiss terrane of our earlier terminology from arc-related volcanoplutonic (greenstone-granite) rocks in the southern part of the Superior province. The GLTZ is interpreted as a paleosuture resulting from continent-continent collision. At most places in the Great Lakes region, the tectonic zone is covered by Proterozoic rocks or Pleistocene glacial deposits, and its position and characteristics previously have been determined mainly by geophysical data. Recent geologic mapping, however, of an exposed segment of the GLTZ in northern Michigan (Sims, 1991a, 1993) has indicated for the first time the kinematics of the structure. Collision along the GLTZ possibly can account for the dextral transpression (Hudleston and others, 1988; Schultz-Ela and Hudleston, 1991) imposed on rocks of greenstone-granite affinity to the north of the GLTZ in the Lake Superior region (Sims and Day, 1992, 1993).

The boundary between gneiss and greenstone-granite terranes was first recognized in Minnesota (Sims and Morey, 1973; Morey and Sims, 1976) from regional geologic relations (fig. 9). This Minnesota segment of the boundary was named the Morris fault (Gibbs and others, 1984; Sims and Day, 1992, p. M7, fig. 4). Differences in rock types, degree of metamorphism, structure, age, and origin indicated that the two basement terranes had vastly different geologic histories and probably had evolved separately. These unique differences were established as the basis for distinguishing the two major terranes. The position of the boundary was determined primarily by regional gravity and magnetic data. Later (Sims, 1980), the boundary was approximately delineated in the western part of Upper Michigan (Sims and others, 1984) and northwestern Wisconsin (Sims, Peterman, and others, 1985), on the east side of the Middle Proterozoic Midcontinent rift system. In a still later report, the boundary was named the Great Lakes tectonic zone, and it was inferred on indirect evidence to extend eastward from the Lake Superior region through the Sudbury structure in Ontario, where it is truncated by the Middle Proterozoic Grenville front tectonic zone (Sims and others, 1980). More recently, the Great Lakes tectonic zone was delineated in the subsurface from western Minnesota into central South Dakota, where it is truncated by the Early Proterozoic Trans-Hudson orogen (Sims and others, 1991). The aggregate length of the GLTZ, as defined in the various published reports, is about 1,400 km.

The boundary between subprovinces (GLTZ) has been variously interpreted as a narrow structural zone (Morey and Sims, 1976) and as a broad zone as much as 50 km wide (Southwick, 1980; Sims and others, 1980; Gibbs and others, 1984), which would include the so-called quiet zone (unit Wvq of pl. 1). Further, because the boundary in the Marenisco-Watersmeet area of Upper Michigan coincides with a broad belt of distinctive tectonism in the younger Early Proterozoic supracrustal rocks, the term GLTZ was used to include the Early Proterozoic (Penokean) deformation as well as the Archean (Sims and others, 1980). This broader usage was continued in a later, detailed publication on the Marenisco-Watersmeet area (Sims and others, 1984). For reasons stated following, however, the GLTZ is now believed from outcrop data to be a relatively narrow Archean structure, characterized by mylonite, that formed in Late Archean time, and although it was locally reactivated in the Early Proterozoic, as it was in the Marenisco-Watersmeet area, it should be considered as an Archean structure.

NATURE OF GREAT LAKES TECTONIC ZONE

The Great Lakes tectonic zone in the exposed area south of Marquette, Mich. (fig. 9), is characterized by a mylonite zone about 2.2 km wide that separates Late Archean greenstone-granite rocks from Archean gneisses; the mylonite has been superposed on previously deformed and metamorphosed dominantly granitoid rocks of the Wawa subprovince and layered gneisses of the Minnesota River Valley subprovince (Sims, 1991a). The presence of mylonite, the approximately 2 km width of the zone, and the juxtaposition of greenstone-granite and gneiss terranes distinguish the GLTZ from other less significant ductile-brittle shear zones and faults within the Superior province rocks of the area.

In the Marquette area, foliation in the mylonite strikes N. 70° W. and dips 75° SW. The strike is about 10°–15° more westward than the shear zone walls (Sims, 1993). A pronounced rodding (stretching) lineation in the mylonite expressed mainly by comminuted and recrystallized quartz and quartz-feldspar aggregates plunges an average of 42° in a S. 43° E. direction. Hinges of tight (sheath) folds plunge subparallel to the stretch lineation.

The attitude of the stretching lineation (line of tectonic transport and X finite strain axis) in the mylonite together with asymmetric structures indicative of sense of movement indicates that collision in the Marquette area was oblique, resulting in dextral-thrust shear along the boundary and northwestward vergence and probable overriding of the Wawa subprovince by the gneiss of the Minnesota River Valley subprovince (Sims, 1991a). This implies southward

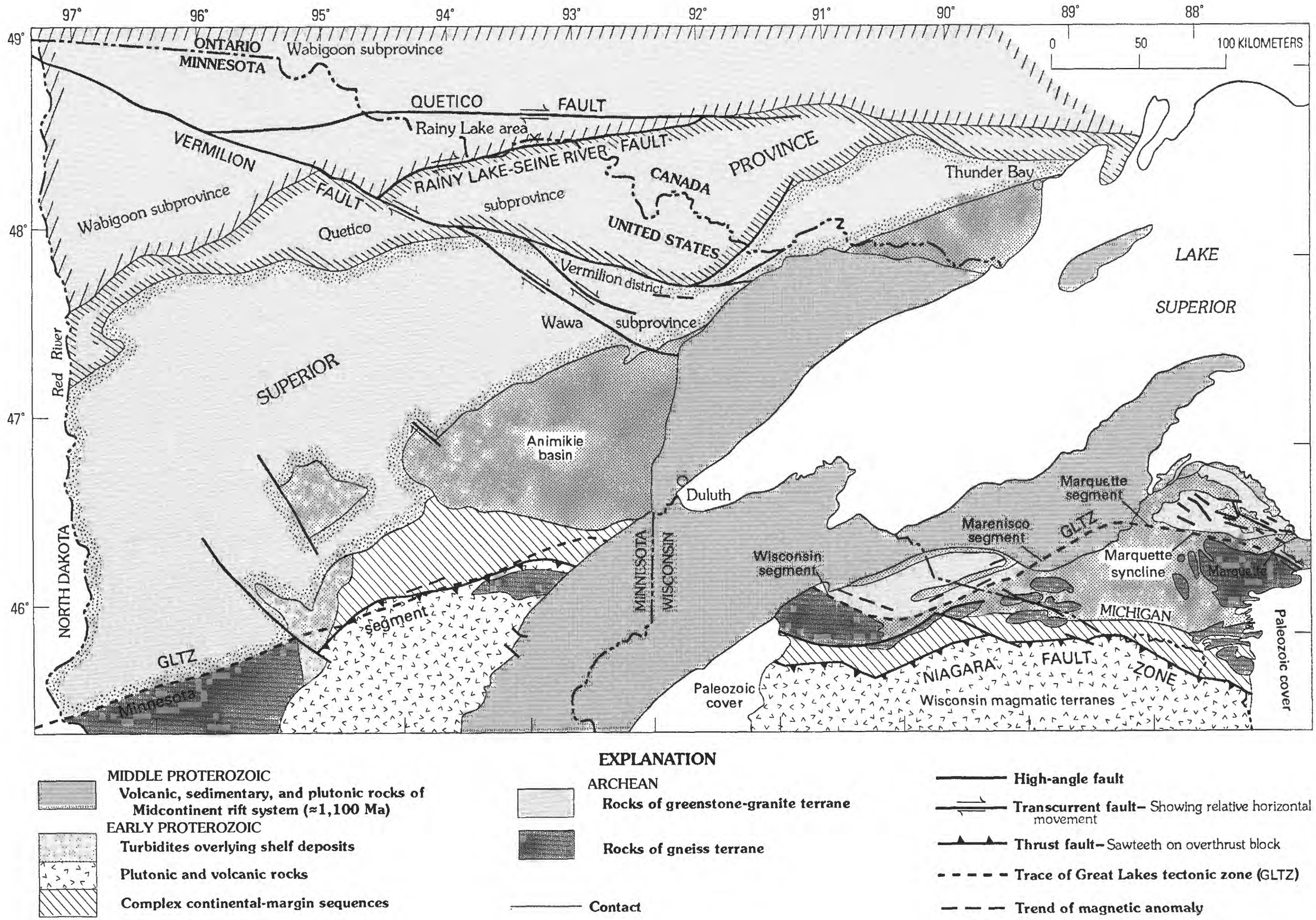


Figure 9. Simplified tectonic elements of Lake Superior region, showing Great Lakes tectonic zone and adjacent Archean terranes. Geology modified from Morey and others, 1982. Greenstone-granite terrane includes three subprovinces; gneiss terrane is Minnesota River Valley subprovince.

subduction of the northern terrane or northward obduction of the gneiss terrane over the northern terrane.

The structural data from the Marquette area indicating the direction of tectonic transport during suturing of the two Archean terranes provide a means for determining the evolution of the GLTZ and the variable trajectory of stress into the Superior province crust (Sims and Day, 1992, 1993). The GLTZ in the Lake Superior region is known from studies of its various segments (fig. 9) to have systematic angular bends that alternately trend northeast and northwest. This zigzag pattern presumably reflects relict irregularities in the margin of the Wawa subprovince crust, which must have been a continental margin before convergence and collision with the gneisses of the Minnesota River Valley subprovince.

The northeast- and northwest-trending segments of the GLTZ have different structural styles (Sims and Day, 1993). Deformation along the northwest-trending segments, as indicated mainly by data from the Marquette segment (fig. 9), caused dextral transpression resulting from oblique collision. Transmittal of this transcurrent shear into the crust north of the GLTZ possibly yielded the widespread, pervasive west-trending foliation, subparallel upright folds, and northwest- to west-trending dextral faults and shear zones in the greenstone-granite terrane, as described by Hudleston and others (1988) and Bauer and Bidwell (1990) from northern Minnesota and Johnson and Bornhorst (1991) from northern Michigan. This deformation followed an earlier episode of recumbent folding. Collision along the northeast segments, on the other hand, was more nearly perpendicular to the continental margin, and produced northeast-trending structures in the crust immediately northwest of the GLTZ, as can be seen from the regional geologic maps (Morey and others, 1982; Sims, 1992). Whether mylonite was developed along the terrane boundaries of this attitude is not definitely known, but shallow drilling in Minnesota did not intersect a major mylonite zone (Southwick, 1980; Southwick, Meyer, and Mills, 1986).

The northwest vergence indicated by field observations in the Marquette segment of the GLTZ suggests that initial collision along the GLTZ (fig. 9) began at promontories on the margin of the Wawa subprovince (Sims and Day, 1993). Oblique compression at these points produced dextral transpression in the crust north of the GLTZ, probably at least as far north as the Quetico fault, a distance from the GLTZ of about 250 km. Such a distance for strain transmittal is consistent with that for the Appalachian-Ouachita orogeny, where Paleozoic carbonate rocks preserve a shortening fabric as much as 800 km distant from the orogenic front (Craddock and van der Pluijm, 1989).

Approximate time constraints on the collision along the GLTZ (Sims, 1991a) are provided by precise U-Pb isotopic analyses of zircon, titanite, and rutile from the Rainy Lake area (Davis and others, 1989), which lies astride the Quetico and Wabigoon subprovinces (fig. 9). Wrench faulting, which is thought to be a direct response of collision along the GLTZ, is bracketed between 2,696 Ma, the age of a clast in a Timiskaming-type conglomerate, and 2,686 Ma, the age of an undeformed granite pluton, and followed earlier regional deformation and metamorphism by 15–20 m.y. Accordingly, the transpressive deformation is tentatively dated at ≈2,690 Ma, although deformation must have been diachronous along the GLTZ and could in part have been younger. Collision along the GLTZ probably was the terminal suturing of the Superior craton (Hoffman, 1989).

INTERPRETATION OF SEISMIC-REFLECTION PROFILING IN MINNESOTA

The prevailing interpretation based on geophysical data obtained in central Minnesota (Gibbs and others, 1984; Southwick and Chandler, 1983) has been that the GLTZ is a north-dipping tectonic feature, which implies southward vergence and probable northward subduction, as postulated by Hoffman (1989) and Card (1990). An alternative interpretation, based on the recent mapping of the 9-km-long segment of the GLTZ in the Marquette, Mich., area (Sims, 1991a; Sims, 1993) and my reinterpretation of the seismic-reflection profiles in Minnesota, is that the GLTZ is a narrow, steep south-dipping structure formed by north-verging deformation resulting from southward subduction (Sims and Day, 1992, 1993).

In central Minnesota, seismic-reflection profiling reveals shallow-dipping seismic reflectors throughout the upper 30 km of the crust (Gibbs and others, 1984, fig. 4), and many of these reflectors dip gently northward (fig. 10). Gibbs and others (1984) and Smithson and others (1986) interpreted these reflectors as probable faults, possibly thrusts. A particularly conspicuous zone of north-dipping reflectors projects to the surface about 5 km north of the position of the GLTZ, as previously determined by Morey and Sims (1976) from analysis of gravity and aeromagnetic data; and this zone of numerous reflectors was interpreted by Gibbs and others (1984) as probably being the GLTZ, the suture between the greenstone-granite terrane (Wawa subprovince) on the north and the gneiss terrane (Minnesota River Valley subprovince) on the south (fig. 10). Other shallow-dipping reflectors both above and below the presumed GLTZ were interpreted as possible imbricate thrust structures related to the suturing; thus in this interpretation, the GLTZ is a moderately wide zone. A detailed aeromagnetic survey augmented by various computer-generated graphic enhancement techniques, by computer-generated mapping of the second vertical derivative of the gravity field, and by shallow test-drilling supports a major north-dipping structure between the greenstone-granite and gneiss terranes in central Minnesota (Southwick and Chandler, 1983).

The alternative interpretation of the Minnesota seismic data (Sims and Day, 1992, 1993) is based on (1) the observation from outcrop that the GLTZ is a steep, relatively

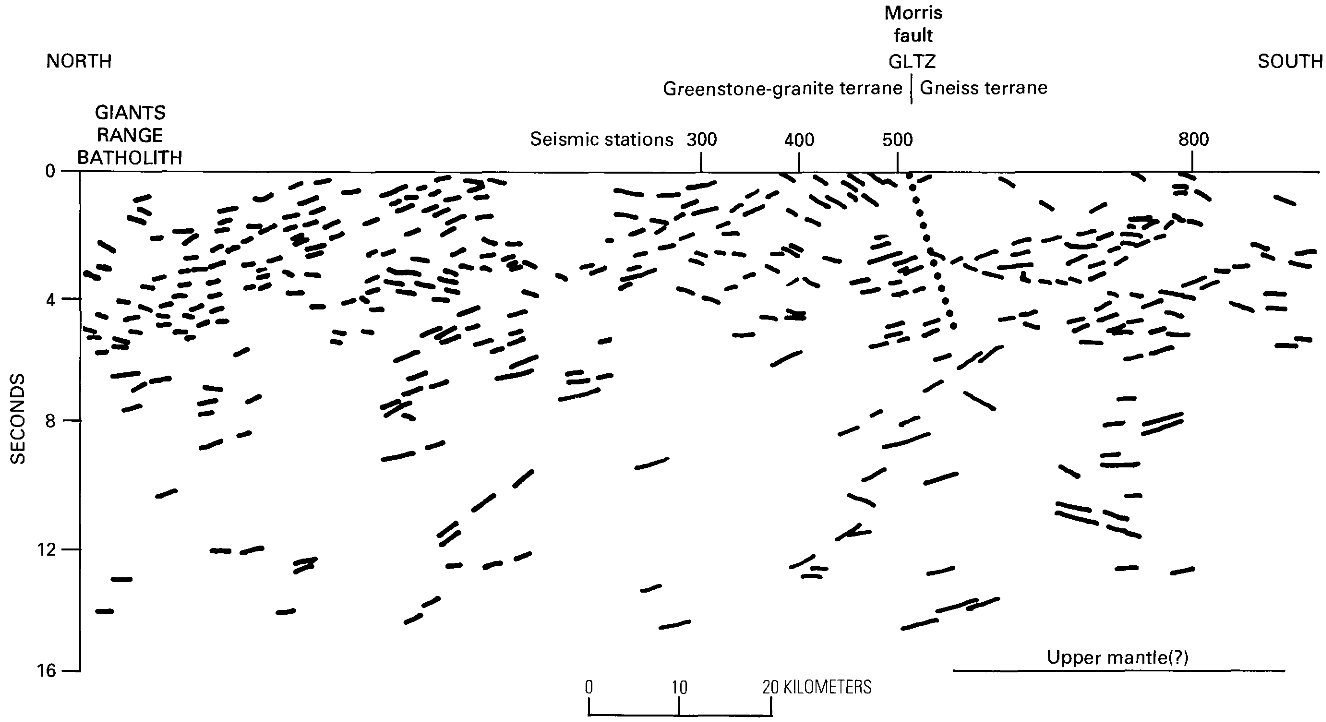


Figure 10. Line drawing of COCORP north-south seismic-reflection profile, Minnesota line 3. These are time sections. Modified from Gibbs and others, 1984, figs. 4, 6. Dots, Morris fault, reinterpreted position of Great Lakes tectonic zone by Sims and Day (1993). Greenstone-granite terrane is now called Wawa subprovince; gneiss terrane is now called Minnesota River Valley subprovince.

narrow, south-dipping high-strain zone, and (2) analogy with seismic-reflection profiles in the greenstone terrane of the Abitibi subprovince to the northeast in Ontario and Quebec (Jackson and others, 1990; Clowes and others, 1992).

In the Abitibi belt, relatively flat seismic reflectors characterize the upper 12 km of greenstone crust (Jackson and others, 1990, fig. 2). These reflectors resemble those in the greenstone crust of Minnesota in being discontinuous and having variably oriented shallow dips. The reflectors in the Abitibi subprovince are truncated and in part offset by major regional, steeply dipping ductile-brittle shear zones, such as the Destor-Porcupine fault, which because of small acoustic impedance contrasts and steep dips are transparent in seismic profiles. The shallow-dipping reflectors are interpreted by Jackson and others (1990) as layering and (or) tectonic high-strain features, perhaps low-angle faults that produced an "out of sequence" stratigraphy revealed by U-Pb geochronology (Corfu and others, 1989).

Sims and Day (1992, 1993) have proposed that the shallow-dipping seismic reflectors in the greenstone crust of central Minnesota probably represent recumbent folds of layered rocks, such as those that characterize the greenstone-granite terranes in outcropping areas in northern Minnesota and Michigan, and perhaps related axial-plane(?) faults. The zone of prominent north-dipping reflectors earlier interpreted as possibly being the GLTZ (Gibbs and others, 1984) could be an uncommonly continuous zone of well-defined layering. Further, Sims and Day interpreted the steep southeast-dipping fault near station 500 on seismic line 3 (fig. 10), delineated by Gibbs and others (1984) from their analysis of the seismic data, as the GLTZ. This structure is transparent in the seismic profiles, like the shear zones in the Abitibi belt, and has the local name, Morris fault (Gibbs and others, 1984, fig. 2). Surface-wave focal plane determinations of a recent earthquake near Morris, Minnesota (Hermann, 1979), indicate a probable southward dip (≈70°) for the Morris fault. This attitude for the structure is consistent with the observed dip of the GLTZ in Michigan (Sims, 1991a, 1993), as well as with northward vergence on the structure. Also, the apparent narrowness of the Morris fault in Minnesota is consistent with the ≈2-km-wide mylonite zone in the exposed segment in Michigan.

REACTIVATION OF TECTONIC ZONE

The northeast-trending segments of the GLTZ, as represented by the Marenisco segment (fig. 9), were strongly reactivated during Early Proterozoic (Penokean) deformation, whereas the northwest-trending segments apparently were little disturbed by Penokean tectonism. In the Marenisco-Watersmeet area, Michigan (Sims and others, 1984), Penokean deformation produced open to tight northeast-trending folds overturned to the northwest in Early

Proterozoic layered rocks. A southeast-dipping axial-plane foliation was superposed on previously deformed rocks of Late Archean age along the south margin of the greenstone-granite terrane (Wawa subprovince). The Early Proterozoic structures are nearly parallel to the Archean structures and to the orientation of the GLTZ. Gneiss doming (Watersmeet dome) accompanied the Early Proterozoic deformation and took place in the interval 1,800–1,750 Ma. All the Archean rocks in the Watersmeet dome were deformed and metamorphosed during the Penokean event and Rb-Sr whole rock and mineral systems in all the rocks were reset at about 1,750 Ma. The vergence of the regional Penokean folds in the Proterozoic rocks indicates a northwest-oriented maximum compressive stress, which is virtually parallel to the direction of tectonic transport during Late Archean suturing.

In contrast, the northwest-oriented segments of the GLTZ show little geologic or isotopic evidence for Early Proterozoic tectonism. Studies of Archean rocks in northwestern Wisconsin, along and adjacent to the Wisconsin segment of the GLTZ (Sims, Peterman, and others, 1985), showed that Late Archean deformation produced a west-northwest-oriented foliation and gently plunging fold axes, mainly to the southeast; a younger deformation, largely brittle faulting, was accompanied by retrogressive metamorphism. The younger deformation took place during Keweenawan rifting at ≈1,050 Ma, as indicated by Rb-Sr isochron age data. Evidence for Early Proterozoic (Penokean) deformation and metamorphism is lacking. In the same way, evidence for superposed Early Proterozoic deformation along the Marquette segment of the GLTZ is absent except for those Archean rocks in and immediately adjacent to the Marquette syncline (fig. 9). The Marquette syncline truncates structures in the Archean basement.

The strong reactivation of Archean rocks along the northeast-trending segments of the GLTZ and the virtual absence of either geologic or isotopic evidence for reactivation along the northwest-trending segments together suggest that the orientation of the GLTZ had a major influence on the local intensity of the Penokean tectonism. Where the GLTZ was oriented about perpendicular to the Early Proterozoic maximum compressive stress, as in the Marenisco segment, the basement was strongly reactivated and Early Proterozoic folds formed parallel to the trend of the GLTZ. Where the GLTZ is oriented approximately parallel to the Early Proterozoic maximum compressive stress, the Archean structure impeded the development of Early Proterozoic folds and related structures.

Some geologic evidence indicates that the GLTZ has been intermittently active in post-Penokean time. In Minnesota, there are poorly documented changes in the thickness of Upper Cretaceous marine sedimentary rocks across the GLTZ, and corresponding changes in the attitude of the sub-Cretaceous unconformity (Shurr, 1980; Dutch, 1981a; Setterholm, 1990). These data suggest mild vertical tectonism in Late Cretaceous to Tertiary time. Finally, the recent seismic history of Minnesota suggests that the GLTZ is a continuing locus of low-level stress release in the continental crust (Mooney, 1979), although some uncertainties exist as to the seismogenic role of the GLTZ itself (Chandler and Morey, 1989). Chandler and Morey (1989) have suggested that the principal epicenters are more likely related to structures, mainly trending northwestward, that are younger than the Morris fault (GLTZ).

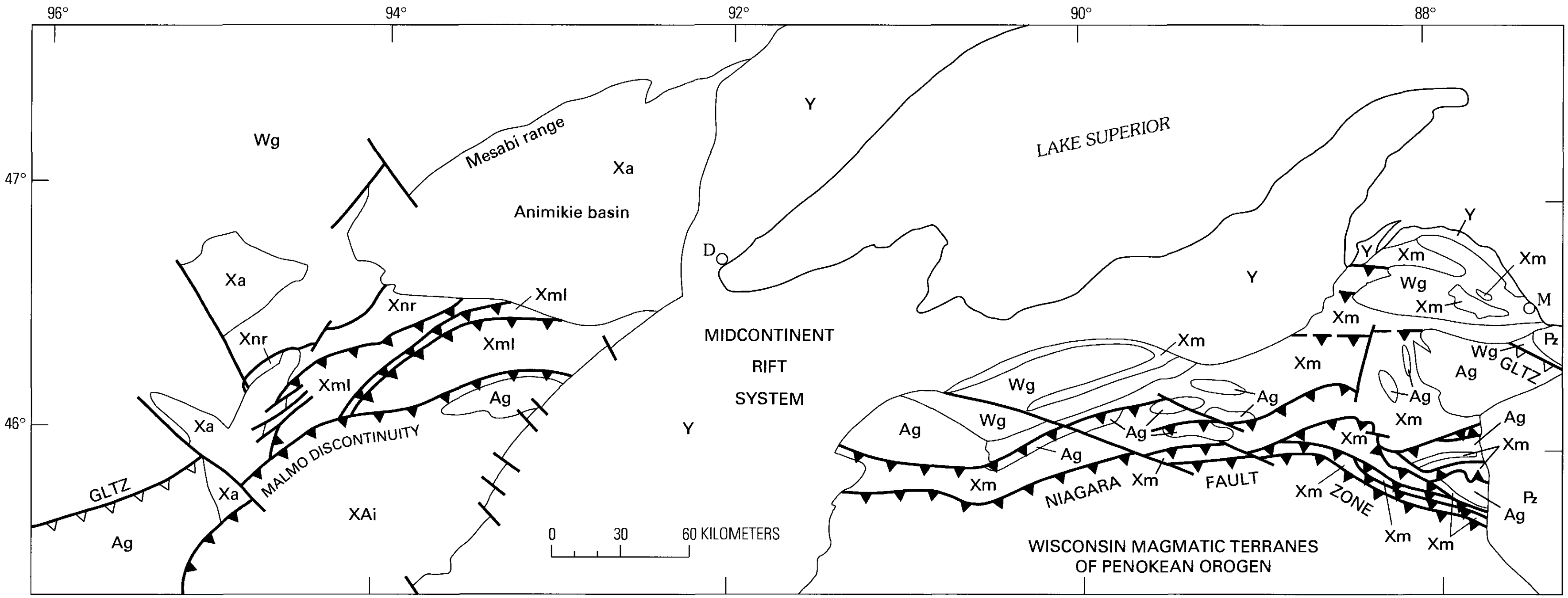
EARLY PROTEROZOIC PENOKEAN OROGEN

By P.K. Sims

The Penokean orogen developed along the south margin of the Archean Superior craton during the Early Proterozoic, and now is known to extend at least from Minnesota eastward across the north shore of Lake Huron to the Grenville front (pls. 1 and 2). Because of past debate about possible correlations of the Early Proterozoic rocks of the Lake Superior region and those of the Lake Huron region, this subject is reviewed briefly in this section.

In the Lake Superior region, the Penokean orogen comprises a deformed, south-facing, passive continental margin assemblage overlying an Archean basement that is juxtaposed with southern magmatic arc terranes termed the Wisconsin magmatic terranes (fig. 11) (Sims and others, 1989). The juncture of the two terranes is the Niagara fault zone in Michigan and Wisconsin and the Malmo discontinuity in Minnesota (Southwick and Morey, 1991), which form the southern limit of Archean crust known to be continuous with the Superior craton. Suturing of the two terranes took place at ≈1.85 Ga. Deformation was renewed in the Wisconsin magmatic terranes about 1.84 Ga, when the Pembine-Wausau and Marshfield terranes were sutured.

The continental margin assemblage in the Lake Superior region consists of the Marquette Range Supergroup in Michigan and Wisconsin (Cannon and Gair, 1970) and the



EXPLANATION

Pz	Paleozoic rocks
Y	Midcontinent rift system (Middle Proterozoic)
	EARLY PROTEROZOIC
Xa	Sedimentary rocks of Animikie Group
Xnr	Sedimentary rocks of North Range Group of Southwick and others (1988)
Xml	Sedimentary and volcanic rocks of Mille Lacs Group
Xm	Sedimentary and volcanic rocks of Marquette Range Supergroup
	EARLY PROTEROZOIC AND ARCHEAN
XAi	Granitoid rocks intruded into complex metamorphic terrane
	ARCHEAN
Wg	Rocks of greenstone-granite terrane (Wawa subprovince)
Ag	Rocks of gneiss terrane (Minnesota River Valley subprovince)

Contact

High-angle fault

Thrust fault of Early Proterozoic age—Dashed where inferred; sawteeth on upper plate

Thrust fault of Archean age—Sawteeth on upper plate

D Duluth

M Marquette

Figure 11. Generalized geology of major part of Early Proterozoic Penokean orogen, Lake Superior region. Compiled by P.K. Sims, 1991. GLTZ, Great Lakes tectonic zone.

Mille Lacs, Animikie, and North Range Groups in Minnesota (Morey, 1983a,b; Southwick and others, 1988). They comprise a lower rifted passive margin sequence overstepped northward by a synorogenic foredeep sequence (Hoffman, 1987; Barovich and others, 1989; Southwick and others, 1988). The Early Proterozoic deformation produced a foreland fold-and-thrust belt that involved Archean basement rocks (Klasner and Sims, 1993; Southwick and Morey, 1991) and the production of gneiss domes (Attoh and Klasner, 1989; Gregg, 1993). The southern, magmatic terranes consist of Early Proterozoic calc-alkaline and tholeiitic volcanic and calc-alkaline plutonic rocks, interpreted as island arc and (or) back-arc basin deposits (Sims and others, 1989).

The Niagara fault zone is a broadly arcuate, convex-northward system of north-verging ductile shears as much as 10 km wide (Sims, 1990b) that contains flattened, steeply dipping rocks having down-dip stretching lineations (Larue and Ueng, 1985; Sedlock and Larue, 1985; Sims and others, 1992). It contains a dismembered ophiolite and has been interpreted as a paleosuture that juxtaposes the magmatic terranes on the south and the continental margin (Larue, 1983; Schulz, 1987; Sims and others, 1989). Southwick and others (1988) have equated the Malmo discontinuity in east-central Minnesota with the Niagara fault zone.

The present consensus that the evolution of the Penokean orogen involved a plate tectonic regime has evolved from several studies during the past decade, and particularly from the evolutionary model of Schulz and others (1993). Reports for the segment in Minnesota include Holst, 1984; Morey and Southwick, 1984; Southwick and others, 1988; Holm and others, 1988; Southwick and Morey, 1991; and Holst, 1991, and reports for the segment in Upper Michigan and Wisconsin include Van Schmus, 1976; Cambray, 1978; Larue, 1983; Sims and Peterman, 1983; LaBerge and others, 1984; Schulz, 1984; Sims, Peterman, and others, 1985; Sims and others, 1987; Klasner and others, 1988; Attoh and Klasner, 1989; Klasner and Sims, 1993; and Gregg, 1993. Earlier models for the Penokean orogeny emphasized intracratonic tectonism as a principal mechanism (Cannon, 1973; Sims, 1976; Morey, 1979; Sims and others, 1981), a logical extension of James' (1954) early model that the Marquette Range Supergroup evolved from a "stable-shelf" to a "geosynclinal" assemblage.

CONTINENTAL MARGIN ASSEMBLAGE

By G.B. Morey

INTRODUCTION

Supracrustal sequences of Early Proterozoic age in the Lake Superior and Lake Huron regions constitute a discontinuous linear foldbelt some 1,300 km long, which extends from eastern Ontario into Minnesota along the south margin of the Superior province of the Canadian Shield (fig. 12). The sequences compose a major part of the so-called "Southern Province" of the Canadian Shield and are referred to here as part of the Penokean orogen. The orogen is transected at both ends of Lake Superior by the Middle Proterozoic Midcontinent rift system and, in northeastern Ontario, by the Late Proterozoic Grenville Front tectonic zone, which largely obliterates the primary features of Early Proterozoic rocks of the Grenville province.

From east to west the supracrustal sequences are the Huron Supergroup and Whitewater Group in the Lake Huron region of Ontario, the Marquette Range Supergroup in northern Michigan and Wisconsin, and the Mille Lacs, Animikie, and North Range Groups in Minnesota. All are composed dominantly of epiclastic rocks, but subordinate volcanic and hypabyssal intrusive rocks are present locally. The Lake Superior region also contains appreciable chemical sedimentary rocks, chiefly iron-formation, whereas the Lake Huron region lacks iron-formation and contains several widespread glaciogenic sequences not recognized in the Lake Superior region. These differences have led to considerable controversy regarding possible correlations between the two regions.

The beds of iron-formation that crop out widely in the Lake Superior region were formerly believed to represent a unique temporal event that could be used to correlate isolated sequences (for example, Van Hise and Leith, 1911). The idea of a principal episode of iron-rich deposition persisted into the 1980's (for example, Morey, 1983b; Morey and Van Schmus, 1988; and especially Morey, 1993). It was first challenged by the structural studies of Southwick and others (1988) in Minnesota, who showed that the Penokean orogen consists of two major components, an allochthonous fold-and-thrust belt on the southeast and one or more tectonic foredeeps on the northwest. The fold-and-thrust belt consists of several discrete structural panels (or terranes) bounded by

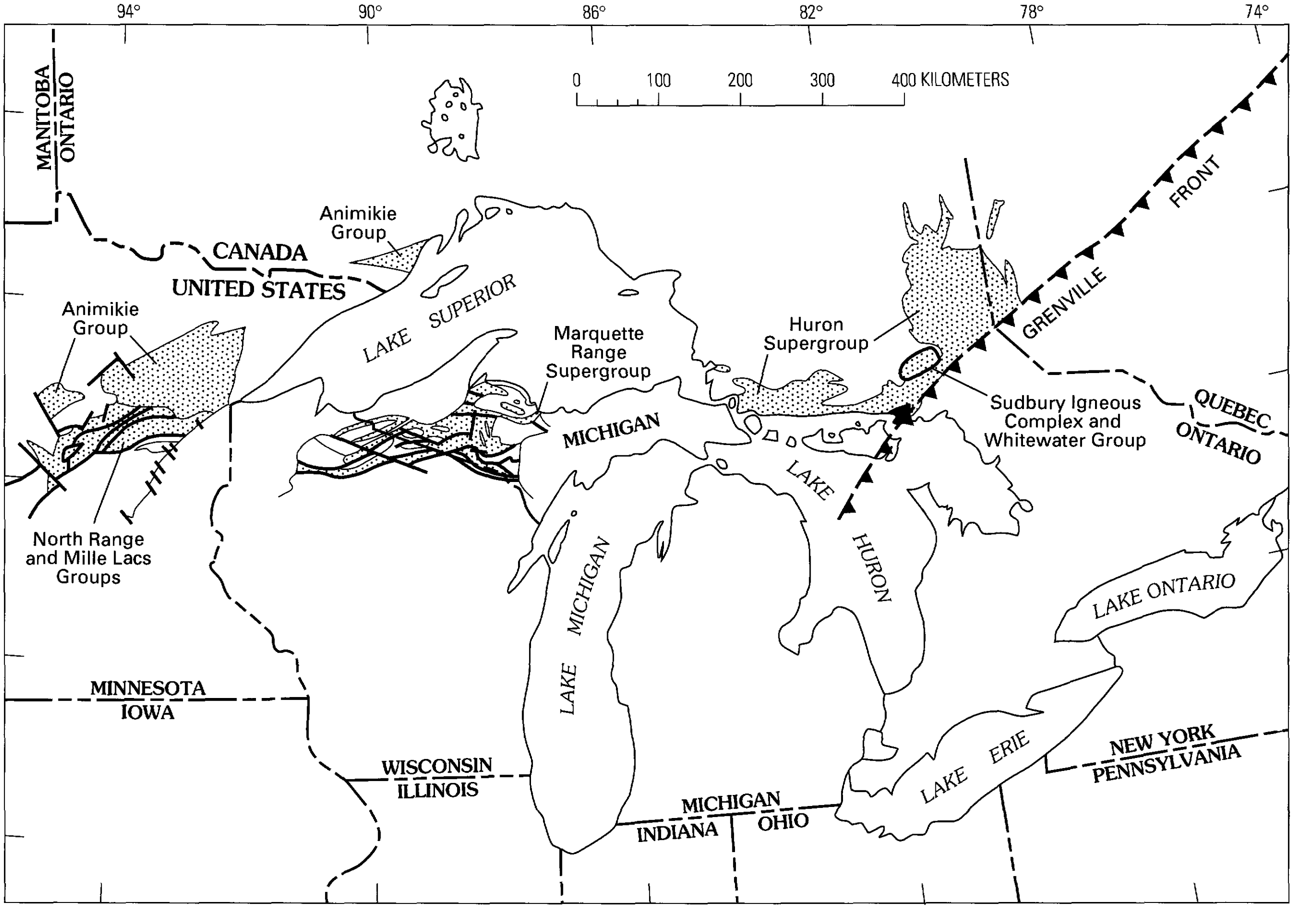


Figure 12. Generalized distribution of Early Proterozoic rocks in Great Lakes region. Stipple, mainly sedimentary rocks; black, mainly volcanic rocks. Dashed line, Grenville front; sawteeth on upper plate. Modified from Young (1983).

discontinuities that have small-scale features consistent with large-scale northwest-verging nappes (Southwick and Morey, 1991). The fold-and-thrust belt flanks a tectonic foredeep which extended to the Mesabi range in northern Minnesota and the Gunflint range in Ontario, the type locality for the well-known Animikie Group. Because the foredeep is filled with strata assigned to the Animikie Group, Southwick and others (1988) have referred to it as the Animikie basin; this usage of "Animikie basin" is considerably more restricted than that used in earlier literature (Morey, 1983b; Morey and Van Schmus, 1988).

Barovich and others (1989) used Sm-Nd techniques to show that the lower part of the Marquette Range Supergroup in Michigan had a northern Archean cratonic provenance, whereas the upper part had a southern Early Proterozoic volcanic provenance. Although the structural details are complicated (Sims, "Structure of continental margin," this volume), the combined studies of Southwick and others (1988) and Barovich and others (1989) show that the Penokean orogen can be generally divided into an early extensional phase, when sedimentary rocks, including iron-formation, were deposited on an evolving continental margin, and a subsequent compressional phase, when these continental margin deposits were folded and metamorphosed and partly overridden by one or more northward-migrating foredeep basins partly filled with iron-formation as well as detritus from a newly formed island arc—the Wisconsin magmatic terranes to the south. Finally, the island arc and the continental margin were sutured together at about 1,850 Ma.

That iron-formation was deposited during both extensional and compressional phases of the Penokean orogeny is of stratigraphic significance and warrants a brief discussion of this new Lake Superior region tectonic paradigm, which has developed mainly over the last 5 years. A brief discussion of the Early Proterozoic rocks in the Lake Huron region also is included, principally because they have had a tectonic history very similar to that of the Lake Superior region and are indeed a part of the Penokean orogen (Young, 1983, 1991; Card, 1989).

LAKE HURON REGION

The Huron Supergroup north of Lake Huron (fig. 13) is one of the best exposed and most accessible Early Proterozoic successions of the world (Young, 1983, 1991) and is

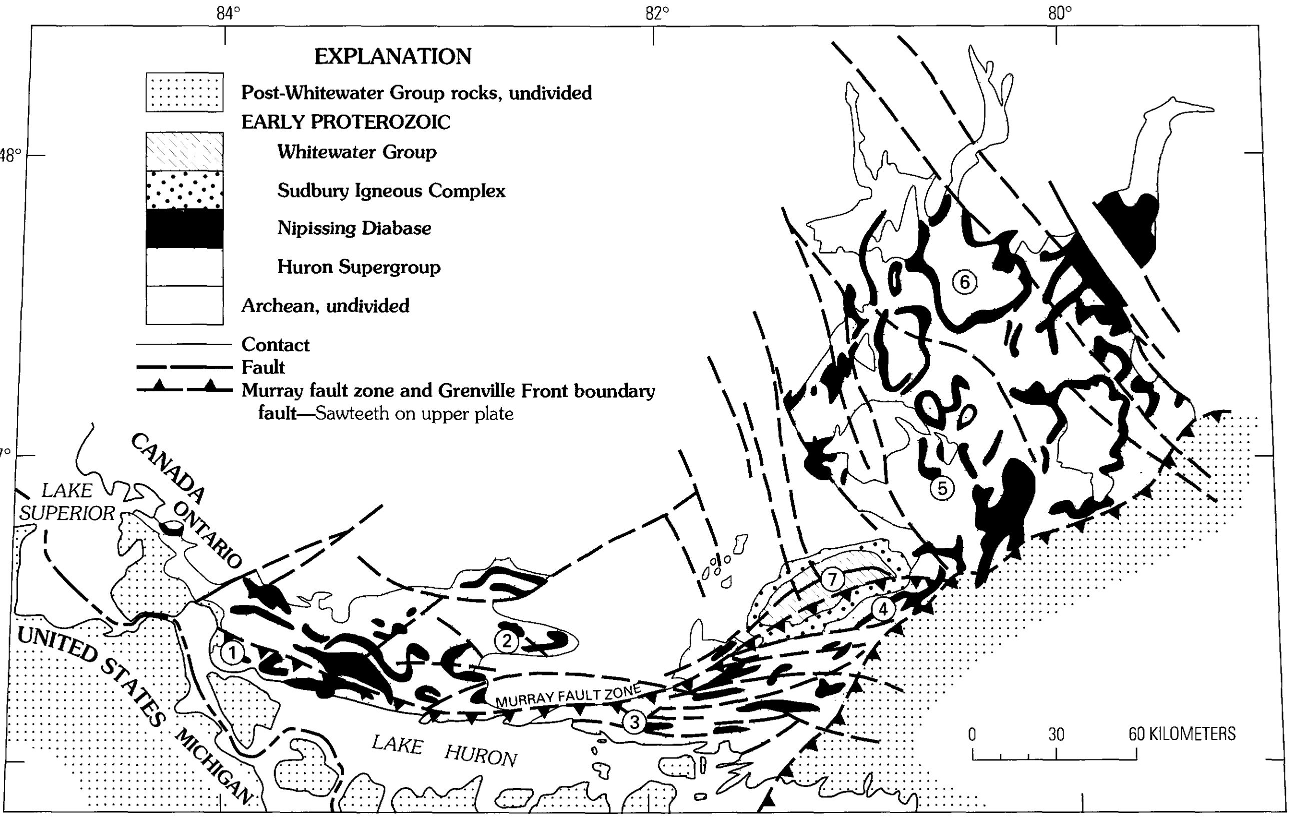


Figure 13. Distribution of Huronian and related Early Proterozoic rocks on the north shore of Lake Huron. Circled number, location of stratigraphic section (fig. 14). Modified from Card and Pattison, 1973; Young, 1983, 1991; Card, 1989 and references therein.

well known for the world-class uranium deposits that it contains. It is of special interest because it contains evidence of fluctuating climatic conditions and includes the world's oldest widespread and well-documented glaciogenic deposits.

The Huron Supergroup varies considerably in thickness, with the thicker parts generally being to the south of the Murray fault zone (fig. 14). Individual groups thin to the north and west of the fault zone, mainly by the attenuation of individual formations, by the wedging out of some units, and by erosion beneath disconformities and unconformities within the succession.

The Huron Supergroup is subdivided into four groups, from oldest to youngest, the Elliot Lake, Hough Lake, Quirke Lake, and Cobalt Groups. The lower three groups are informally termed "lower Huronian," and the Cobalt is informally designated as "upper Huronian." An angular unconformity of regional extent separates lower and upper Huronian strata.

Lower Huronian sedimentation was accompanied by early volcanism in depositional basins bounded by major listric faults displaying south-side-down movement. The volcanic rocks are fissure eruptions related to deep-penetrating faults and include both mafic and felsic lavas and pyroclastic rocks (Card, 1978). One felsic volcanic unit from the upper part of the sequence yielded a U-Pb zircon age of 2,450±25 Ma (Krogh and others, 1984). The volcanic rocks are intercalated with units of graywacke, aluminous pelite, and sulfide-rich strata, and are intruded by layered gabbro-anorthosite bodies. These layered bodies from the lower part of the sequence yield U-Pb zircon ages of 2,491±5 Ma and 2,480±10 Ma (Krogh and others, 1984). The lower part of the Elliot Lake Group is intruded by the Murray and Creighton Granites, which yield U-Pb zircon ages of 2,388±20 Ma (Krogh and others, 1984) and 2,333±33 Ma (Frarey and others, 1982), respectively.

The basal volcanic sequence is overlain by the Matinenda Formation, host to the world-famous Elliot Lake-type uranium deposits. The uranium occurs as disseminated uraninite and brannerite in lenses of pyritic quartz-pebble conglomerate enclosed in crossbedded arkosic sandstone. The Matinenda is overlain by the McKim Formation, a turbiditic sequence made up of graywacke, laminated siltstone, aluminous pelite, and minor quartz arenite.

The overlying two groups of the lower Huronian, the Hough Lake and Quirke Lake, represent two sedimentary megacycles, each consisting of a lower conglomeratic suite, the Ramsay Lake and Bruce Formations, respectively, a middle siltstone-graywacke unit, the Pecors and Espanola Formations, and an upper crossbedded sandstone unit, the Mississagi and Serpent Formations. Most of the

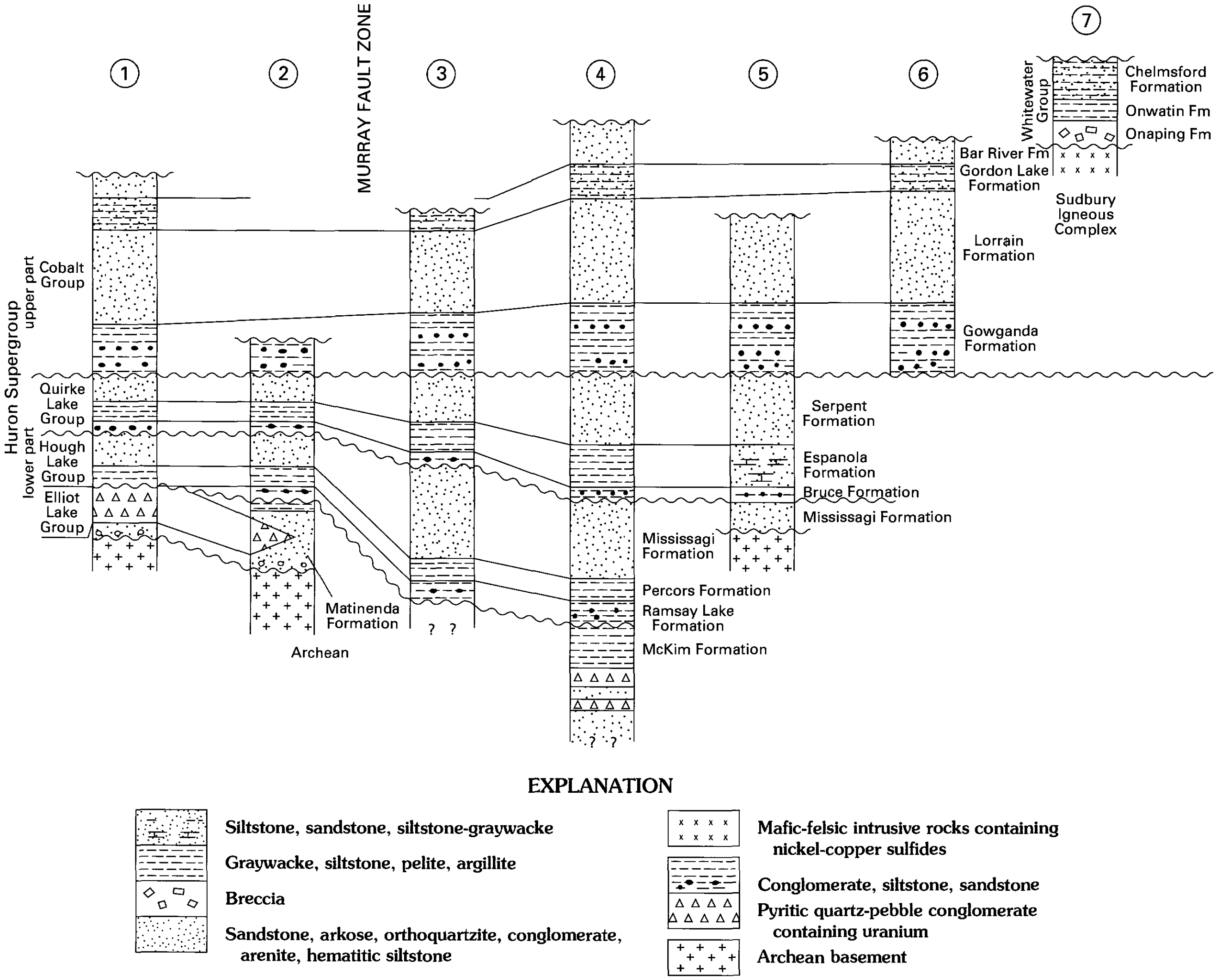


Figure 14. Lithostratigraphic sections and correlation of Huron Supergroup and related Early Proterozoic rocks on the north shore of Lake Huron. Locations of numbered stratigraphic sections shown in figure 13. Wavy line, unconformity; query, lowermost rocks uncertain. At locality 2, volcanic rocks underlie Matinenda Formation. Modified from Frarey and Roscoe, 1970; Card and others, 1984; Dressler, 1984.

conglomerate is matrix-supported paraconglomerate that may have formed by glacial processes. The middle siltstone-graywacke units were deposited by turbidity currents, and the lower part of the Espanola Formation contains beds of limestone, dolomite, and calcareous siltstone that display shallow-water features such as mudcracks and stromatolites. The overlying sandstone units are arkosic and also were deposited in fluvial to shallow marine environments.

The upper Huronian Cobalt Group comprises (1) a lower platform sequence of glaciogenic conglomerate, siltstone, and sandstone, the Gowganda Formation; (2) a thick sequence of arkose, orthoquartzite, and conglomerate, the Lorrain Formation; (3) a thin unit of varicolored siltstone and sandstone, the Gordon Lake Formation; and (4) an uppermost sequence of quartz arenite and hematitic siltstone, the Bar River Formation. The lower part of the Gowganda contains layers of polymictic conglomerate (diamictite) separated by units of laminated siltstone (varvite) deposited during several glacial advances from the north. The upper part of the Gowganda consists of coarsening-upward cycles of siltstone and sandstone that typically display large-scale slumping phenomena, possibly associated with prograding deltas. The Lorrain Formation was deposited in fluvial to shallow marine environments, whereas the Gordon Lake Formation has features indicative of shallow subtidal to storm-dominated marine shelf environments. The Bar River Formation also probably was deposited in a transitional shallow tidal marine to beach to eolian environment.

The Huron Supergroup was deformed prior to the emplacement of tholeiitic gabbro sills and dikes, collectively referred to as the Nipissing Diabase. Nipissing Diabase from the Gowganda area yields a U-Pb baddeleyite age of 2,219±4

Ma (Corfu and Andrews, 1986), but according to Buchan and Card (1985) paleomagnetic data provide evidence for more than one Nipissing event. Intrusion of the Nipissing Diabase was followed by renewed deformation and regional metamorphism (Card, 1978) at about 1,900±50 Ma (Fairbairn and others, 1969).

A considerably different Early Proterozoic succession, the Whitewater Group, crops out within the Sudbury basin, an elliptical depression that straddles the Archean-Huronian unconformity north of Sudbury (fig. 13). Stratigraphic relationships between the Huron Supergroup and the Whitewater Group are problematic, as are the nature and origin of the Sudbury basin itself.

The Sudbury basin is outlined by a composite mafic-felsic intrusion called the Sudbury Igneous Complex. World-class massive and disseminated nickel-copper sulfides occur at the base of the complex and in dike-like intrusions termed "offsets," which extend outward from the main body into footwall rocks of the Huron Supergroup (Pye and others, 1984, and references therein).

The Sudbury Igneous Complex is one of the most precisely dated igneous bodies in the world. A variety of noritic rocks yield U-Pb zircon ages of 1,850±1 Ma, whereas a granophyric component yields a U-Pb baddeleyite age of 1,850.5±0.3 Ma (Krogh and others, 1984). Basal rocks of the Whitewater Group are cut by 1,849 Ma granophyric intrusions, and thus the complex and the basal part of the Whitewater Group are approximately the same age.

The Whitewater Group is divided into three formations (fig. 14): a lower conglomeratic unit of enigmatic origin, the Onaping Formation; a middle pelitic unit, the Onwatin Formation; and an upper graywacke unit, the Chelmsford Formation. The Onaping Formation was described by Muir and Peredery (1984) as consisting of an unsorted and poorly stratified breccia with clasts derived from both the Archean Superior province and the Huron Supergroup. The clasts are set in a matrix of devitrified glass shards and mineral and rock fragments. The Onaping has been interpreted as either a fallback breccia from meteorite impact or the product of explosive volcanism (Muir and Peredery, 1984).

The Onwatin Formation gradationally overlies the Onaping Formation and consists mainly of carbonaceous and pyritic argillite, siltstone, and minor graywacke. A sulfide-bearing carbonate unit, typically 40 m thick, occurs near the base of the formation. The Onwatin is a pelagic deposit formed in a restricted basin with anoxic bottom waters (Rousell, 1984). The carbonate unit is a sedimentary-exhalative deposit, whereas the graywacke beds were deposited by turbidity currents. The conformably and gradationally overlying Chelmsford Formation consists mainly of thick-bedded graywacke with well-developed Bouma beds and other structures indicative of turbidite deposition (Rousell, 1984). The formation represents a proximal sequence deposited by currents that flowed to the southwest, down a paleoslope that was parallel to the present long axis of the Sudbury basin.

The Onwatin and Chelmsford Formations have been interpreted as the local filling of a meteorite impact basin, but Young (1983) has argued that the Chelmsford is the remnant of a flysch deposit once much thicker and more extensive than the strata now preserved in the Sudbury basin.

Rock units of the Sudbury basin, notably the offset dikes of the Sudbury Igneous Complex, cut the axial surfaces of earlier-formed folds in the Huron Supergroup and were themselves deformed during an event that also produced greenschist-grade assemblages in the Whitewater Group. The deformation produced a series of upright to northward-overturned folds having a well-developed axial-plane cleavage. These folds, as well as the Sudbury Igneous Complex, are cut by a series of south-dipping thrusts with displacements of several kilometers. Card (1989) suggested that the deformation occurred late in the Penokean orogeny. If so, the Sudbury event and the Penokean orogeny are closely related in time.

NORTHERN WISCONSIN AND MICHIGAN

The Early Proterozoic succession in northern Wisconsin and Michigan (fig. 15) has yielded major quantities of iron ore from four mining districts—the Gogebic (loc. 1, fig. 15), Marquette (loc. 5, fig. 15), Iron River–Crystal Falls (loc. 7, fig. 15), and Menominee (loc. 8, fig. 15) ranges. Lesser quantities were obtained from the Felch district (loc. 6, fig. 15). Only the Marquette range is still an iron-ore producer.

The iron-bearing and associated epiclastic rocks and intercalated flood basalts are assigned to the Marquette Range Supergroup (Cannon and Gair, 1970), which traditionally is subdivided into four groups, from oldest to youngest, the Chocolay, Menominee, Baraga, and Paint River Groups (fig. 16).

The oldest recognized units in the Chocolay Group in Michigan include deposits of possible glacial origin variously referred to as the Enchantment Lake and Fern Creek Formations. A glacial origin was questioned by Larue (1981a) and Mattson and Cambray (1983), who suggested instead that they formed by fanglomeratic or mass-wasting processes in a fault-controlled basin or basins. More recently, Sims (1991a) has suggested that the Reany Creek Formation, formerly included in the Chocolay Group, is a Late Archean unit of Timiskaming type, which now are commonly interpreted as forming in Archean analogs to modern pull-apart basins. Nonetheless, the Fern Creek Formation is still considered the best example of a Precambrian glaciogenic deposit in the Lake Superior region (Gair, 1981).

An unconformity of uncertain duration separates the glaciogenic deposits from the remainder of the Chocolay Group. Above that unconformity, sedimentation occurred over a wide area in a relatively stable tectonic environment. Sedimentation started with a widespread, mineralogically mature quartz arenite—the Mesnard, Sunday, and Sturgeon

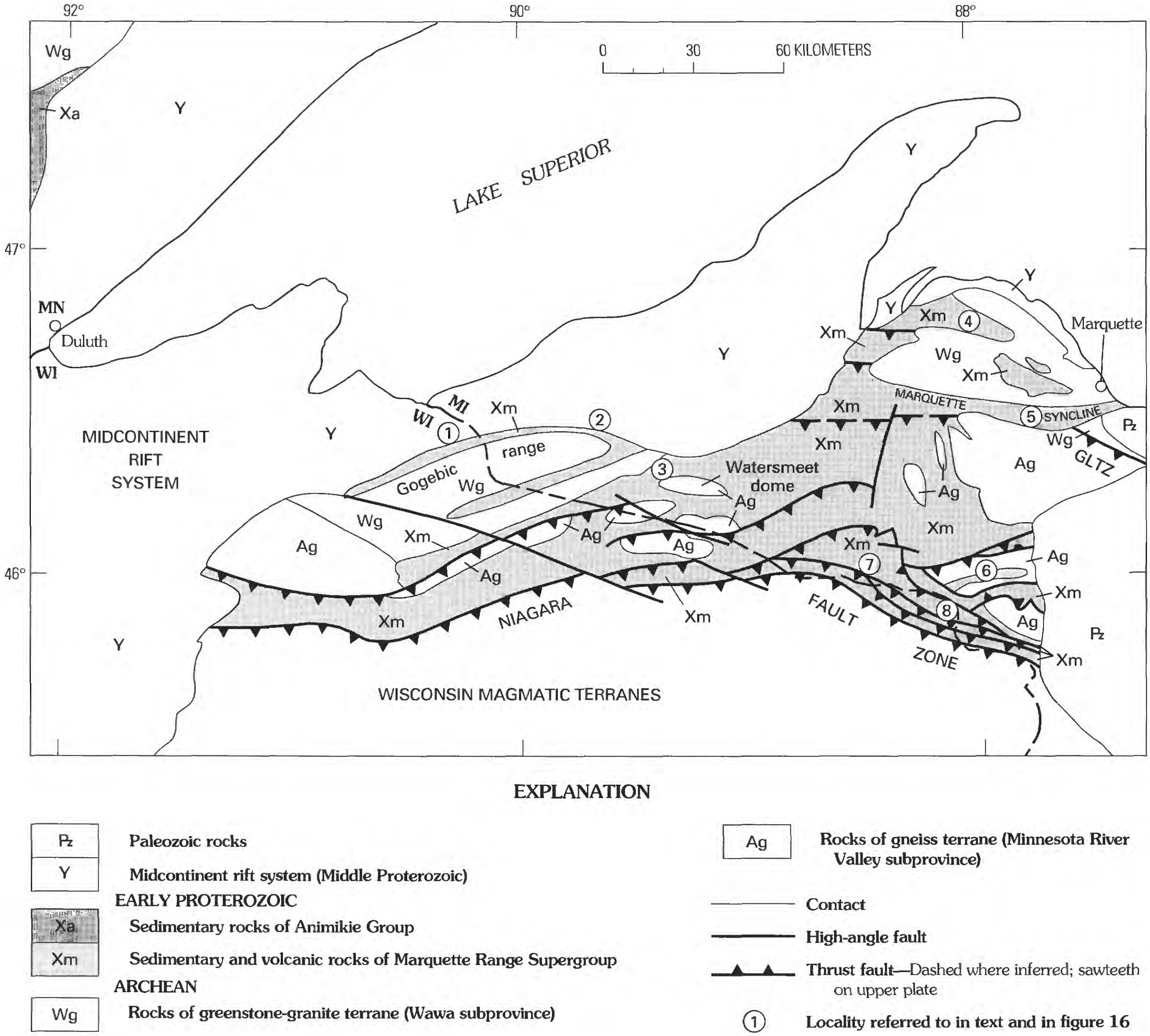


Figure 15. Distribution of the Early Proterozoic Marquette Range Supergroup in northern Wisconsin and Michigan. Circled number, location of stratigraphic section (fig. 16). GLTZ, Great Lakes tectonic zone.

Quartzites—and continued with thick beds of dolomite containing stromatolitic structures—the Kona Dolomite, Randville Dolomite, and Bad River Dolomite. The dolomitic strata are topped by an unconformity everywhere except in the eastern part of the Marquette range (loc. 5, figs. 15, 16), where the dolomite is overlain by the Wewe Slate, a unit of muddy terrigenous, and possibly partly volcanic derivation.

Much of the Chocolay Group was removed by erosion prior to deposition of the Menominee and Baraga Groups. In many places, the Menominee and Baraga Groups are separated by a low-angle unconformity, but in other places the two are interlayered (Cannon, 1986).

Understanding of the Menominee Group stratigraphy has changed since the summary report by Morey (1993) was written. Sedimentation began with regional accumulation of quartz arenite assigned to the Ajibik Quartzite or Felch Formation (loc. 6, figs. 15, 16). On the Marquette range, the Ajibik is overlain by the Siamo Slate, a sequence of intercalated argillite, quartz arenite, and detritus-choked iron-formation. The clastic rocks are overlain by thick deposits of iron-rich strata—the Negaunee Iron-formation, Vulcan Iron-formation, or Amasa Formation—formed for the most part in isolated fault-bounded second-order troughs that dissect the larger subsiding shelf (Larue, 1981b). Iron-formation sedimentation was accompanied locally by subaqueous emplacement of basalt and associated volcanogenic rocks, and lesser amounts of interbedded felsic volcanic rocks, iron-rich strata, and conglomerate. The dominantly volcanic

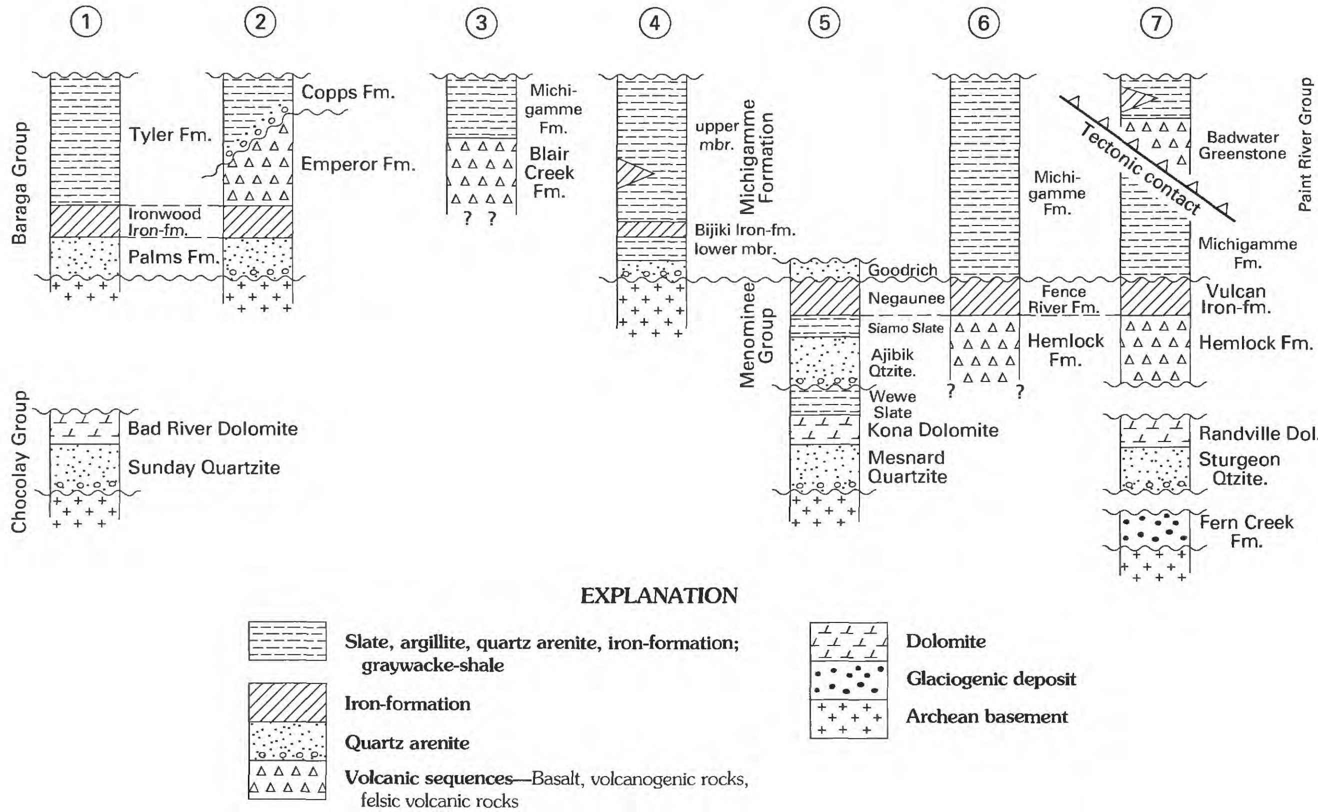


Figure 16. Lithostratigraphic correlations of Early Proterozoic rocks of Marquette Range Supergroup in northern Wisconsin and Michigan. Locations of numbered stratigraphic sections shown in figure 15. Modified from Morey and Van Schmus, 1988, and Morey and Southwick, 1991.

sequences include the Hemlock Formation, which has yielded a U-Pb zircon age of 1,910±10 Ma (Van Schmus and Bickford, 1981). The Badwater Greenstone, previously assigned to the Baraga Group (Cannon, 1986) in the Iron River–Crystal Falls and Menominee ranges, is now correlated with the Hemlock Formation that is assigned to the Menominee Group (Sims, 1990a).

Sedimentation of the Baraga Group in Michigan started on a surface of considerable relief. In the Baraga basin (loc. 4, fig. 15), the group consists of an unnamed quartzite overlying Archean basement (Klasner and others, 1989). The quartzite grades transitionally upward through a thin sequence of lean, cherty iron-formation into graywacke and shale of the Michigamme Formation. The iron-rich strata contain pebble-size clasts of apatite (Mancuso and others, 1975), which yield an essentially concordant Pb-Pb age of 1,929±17 Ma (Zartman, 1987, cited in Klasner and others, 1989). In the Marquette syncline, Baraga sedimentation started with the Goodrich Quartzite, which is overlain by the Michigamme Formation (loc. 4, fig. 15), a thick, turbiditic graywacke-shale sequence. The Michigamme includes several units of iron-formation, some of which, such as the Greenwood and Bijiki Iron-formation Members, are thick enough to be mapped separately. The Michigamme also contains thick volcanic packages, such as the Clarksburg Volcanics Member. On the Iron River–Crystal Falls range (loc. 7, fig. 15), Goodrich Quartzite passes transitionally upward into iron-formation and ferruginous slate—Fence River Formation—which passes into the Michigamme Formation. On the Menominee range (loc. 8, fig. 15), the Baraga Group is similarly dominated by the Michigamme Formation.

The Paint River Group, which is confined to the Iron River–Crystal Falls district (loc. 7, fig. 15), has an uncertain stratigraphic position. It was believed to conformably overlie the Baraga Group (Cannon, 1986, among others). However, geophysical data led Cambray (1987) to speculate that a major discontinuity separates the two groups. Subsequent mapping by Sims (1990a) and colleagues has established that the discontinuity is a thrust fault, and Sims has suggested that the Paint River Group is stratigraphically equivalent to the upper part of the Baraga Group, approximately correlative with the Michigamme Formation.

Stratigraphic relationships on the western Gogebic range (loc. 1, fig. 15) in northern Wisconsin are relatively straightforward, but the existing stratigraphic nomenclature used to describe them (for example, Morey and Van Schmus, 1988) is unnecessarily complex. Tilted and eroded rocks of the Chocolay Group are overlain unconformably by a three-fold sequence of rocks: a lower quartz arenite, the Palms Quartzite; a middle iron-rich unit, the Ironwood Iron-formation; and an upper thick turbiditic graywacke-shale sequence, the Tyler Formation. The Palms-Ironwood contact is gradational over several meters, whereas the Ironwood-Tyler contact is gradational over several tens of meters. Tradition has placed the boundary between the Menominee and Baraga Groups at the Ironwood-Tyler contact mainly to emphasize an Ironwood-Negaunee correlation in the Menominee Group on the one hand and a Tyler-Michigamme correlation in the Baraga Group on the other hand. Because the Palms-Ironwood-Tyler sequence is one continuous package in the central and western parts of the Gogebic range, it is more reasonable to assign the entire sequence to the Baraga Group. In this scheme, the Ironwood is correlated with the Bijiki Iron-formation Member of the Michigamme Formation, as shown in figure 16. Similarly, the Palms can be correlated with the Goodrich Quartzite of the Marquette district and the lower member of the Michigamme Formation in the Baraga basin.

Stratigraphic relationships are more complex at the east end of the Gogebic range (loc. 2, fig. 16) near the Wisconsin-Michigan boundary. There a thin unit of quartz arenite is overlain by beds of iron-formation intercalated with a thick sequence of subaqueous mafic and felsic rocks and coeval hypabyssal intrusions. Traditionally, the quartz arenite has been correlated with the Palms Quartzite and the iron-formation has been correlated with the Ironwood Iron-formation (Van Hise and Leith, 1911). The igneous package is assigned to the Emperor Volcanic Complex (Sims and others, 1984). The volcanic rocks are bimodal, continental rift-related tholeiitic basalt and rhyolite that were deposited under relatively shallow water conditions (Licht and Flood, 1992) in a half-graben basin that deepened to the east (LaBerge, 1992). An unconformity, possibly angular, separates the Emperor Volcanic Complex from sedimentary strata assigned to the Copps Formation. Much of the Copps is a graywacke-shale sequence that resembles and possibly can be correlated with the Tyler Formation. However, the lower part of the Copps is a ferruginous conglomerate and quartzite that resembles the Goodrich Quartzite of the Marquette district.

The correlation scheme for the Gogebic range, outlined previously, is based on the traditional assumption that the iron-formation intercalated with the Emperor volcanic rocks is correlative with the Ironwood Iron-formation. In that scheme, the Emperor-Copps unconformity would pass westward into the contact between the Ironwood and Tyler Formations. An understanding of the stratigraphic relationships clearly must be preceded by careful mapping of the east end of the Gogebic range. Nonetheless, I think that it is significant that the Emperor Volcanic Complex has stratigraphic, structural, and geochemical attributes similar to those that typify volcanic sequences in the Menominee Group in other parts of northern Michigan. Therefore, the iron-formation intercalated with the Emperor is possibly correlative with the Negaunee–Fence River package in other parts of northern Michigan. Thus, the Emperor-Copps unconformity may correspond to the unconformity between the Baraga and Menominee Groups in other parts of Michigan.

MILLE LACS, NORTH RANGE, AND ANIMIKIE GROUPS, MINNESOTA AND ONTARIO

Supracrustal rocks of Early Proterozoic age underlie a broad area in east-central and northern Minnesota (fig. 17); correlative rocks also underlie a much smaller area in far northeastern Minnesota and adjoining parts of Ontario north of Lake Superior. Except for those of the Gunflint range in northeastern Minnesota and Ontario and along the Mesabi range in northern Minnesota, the rocks are very poorly exposed, and investigation has relied heavily on geophysical techniques and drilling (Southwick and others, 1988).

Prior to 1972, the entire Early Proterozoic succession in Minnesota was assigned to the Animikie Group, probably best known for the huge iron ore deposits of the Biwabik Iron-formation on the Mesabi range. On the Mesabi and Gunflint ranges, the Animikie Group unconformably overlies Archean basement. On the Cuyuna range and environs in east-central Minnesota, however, Marsden (1972) recognized an older Early Proterozoic sequence. Subsequently, Morey (1978) subdivided the older sequence into several formations that he collectively assigned to a new unit, named the Mille Lacs Group.

The stratigraphic interpretations of Morey (1978) relied greatly on the concept of a single regional iron-formation, and were developed in terms of pre-plate tectonic concepts. The new plate tectonic interpretation (Southwick and others, 1988) requires that existing stratigraphic nomenclature be revised (fig. 18). In general, rocks of the fold-and-thrust belt can still be assigned to the Mille Lacs Group. However, the belt south of the Serpent Lake discontinuity (fig. 17) is broken into at least four fault-bounded structural terranes, and the rocks cannot be correlated from one terrane to another. Thus, the Mille Lacs Group no longer constitutes the ordered, cohesive stratigraphic sequence envisioned by Morey (1978). Very little is known about the Mille Lacs Group north of the Serpent Lake discontinuity. Sparse drilling suggests that it unconformably overlies Archean basement and consists dominantly of quartz arenite, siltstone, and shale topped by a mappable interval of limestone and dolomite named the Trout Lake Formation (Marsden, 1972).

In the Glen Township–Moose Lake structural terrane (loc. 6, fig. 17), quartz arenite of the Denham Formation is overlain by the Glen Township Formation of Morey (1978), a heterogeneous mixture of pyrite- and pyrrhotite-rich carbonaceous slate, carbonate- and silicate-facies iron-formation, and volcanic rocks of mafic composition (Eldougdoug, 1984). The volcanic rocks yield a Sm-Nd isochron age of 2,197±39 Ma and a Rb-Sr whole rock isochron age of

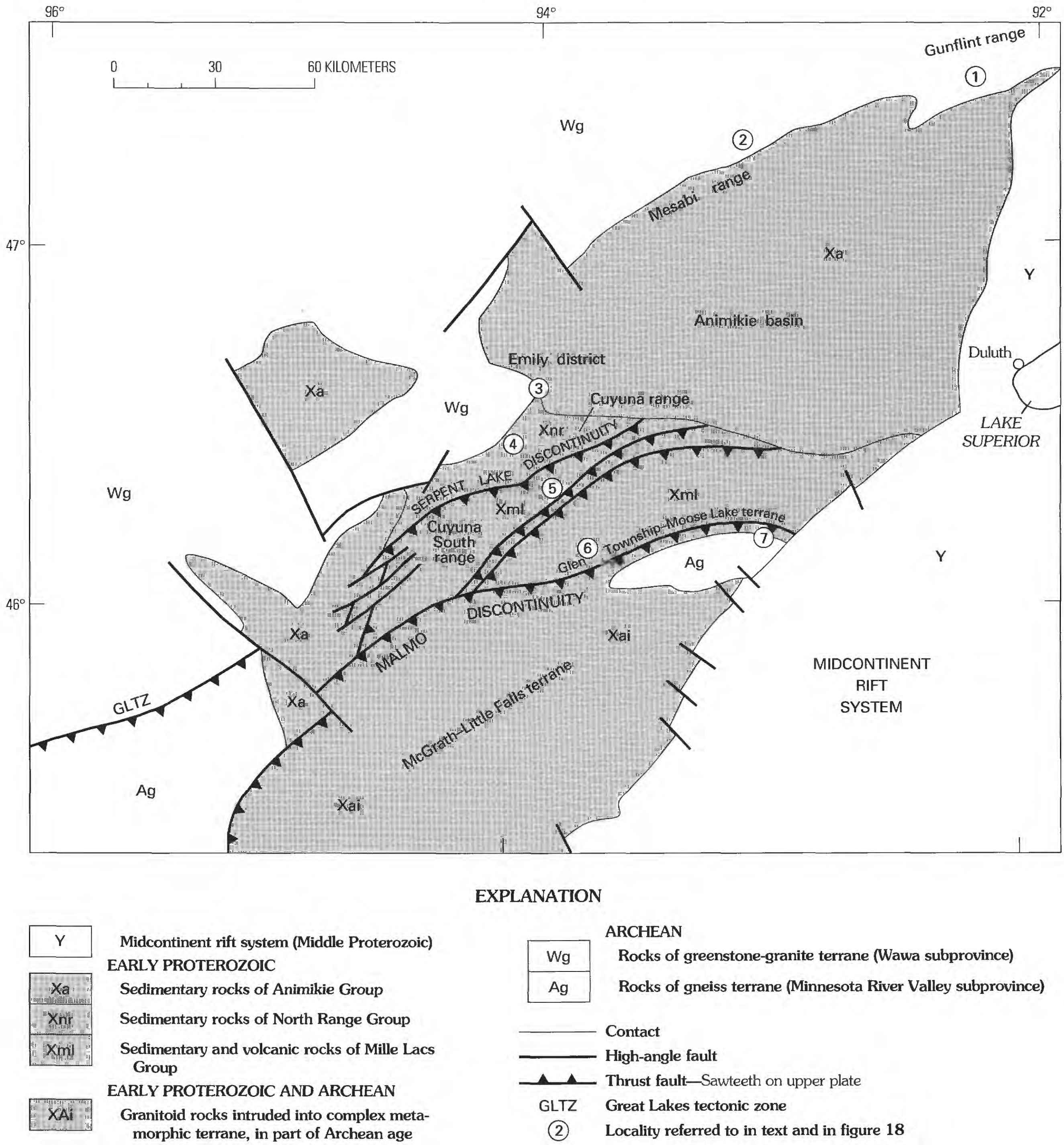


Figure 17. Distribution of Early Proterozoic stratified rocks in the Penokean orogen of east-central Minnesota.

1,746±86 Ma (Beck, 1988). The latter is similar to a Rb-Sr whole rock isochron age of 1,738±16 Ma (Keighin and others, 1972) obtained from a thick graywacke-shale sequence that crops out sparingly in the vicinity of Moose Lake at the east end of the terrane.

Archean basement reappears in the McGrath–Little Falls terrane south of the Malmo discontinuity (loc. 7, fig. 17), where it consists of the McGrath Gneiss and Hillman Migmatite of Morey (1978). The McGrath is overlain unconformably by the Denham Formation of Morey (1978), an Early Proterozoic unit of quartz arenite and quartz-rich siltstone. The lower part of the Denham also contains layers of fluvially deposited conglomerate, beds of oxide-facies iron-formation, and volcanic rocks of mafic to intermediate composition. The upper part contains beds of limestone and dolomite. The McGrath is cut by a tectonized granite gneiss

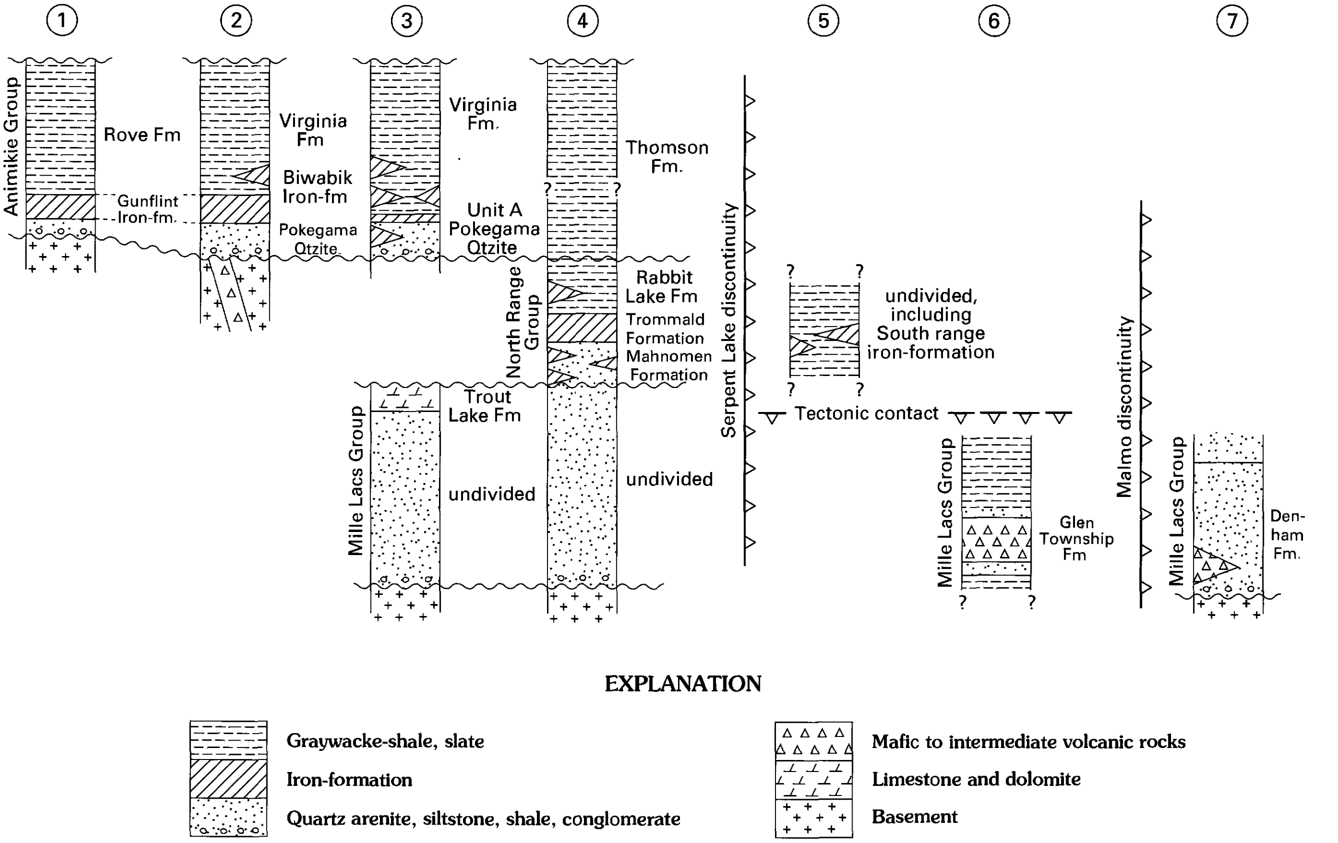


Figure 18. Lithostratigraphic correlations of Early Proterozoic stratified rocks of the Penokean orogen, east-central Minnesota. Locations of numbered stratigraphic sections shown in figure 17. Wavy line, angular unconformity; query, extent of unit uncertain. Modified from Southwick and Morey (1991).

that yields a U-Pb zircon age of 1,982±5 Ma (Goldich and Fischer, 1986), and the Hillman is cut by two syntectonic plutons that yield U-Pb zircon ages of 1,869±5 Ma and 1,868±3 Ma (Goldich and Fischer, 1986). Biotite and hornblende from the McGrath also give similar K-Ar and Rb-Sr mineral ages, in the range of 1,740-1,770 Ma (Keighin and others, 1972).

The North Range Group of Southwick and others (1988) (loc. 4, fig. 17) forms a coherent stratigraphic package between twice-folded rocks of the Mille Lacs Group and once-folded rocks of the Animikie Group. The North Range Group was described by Schmidt (1963), who recognized a lower siltstone-shale sequence, the Mahnomen Formation; an intermediate and dominantly iron rich unit, the Trommald Formation; and an upper graywacke-shale sequence, the Rabbit Lake Formation, together with several thin lenses and layers of iron-formation. Subsequent studies (Morey and Southwick, 1993) show that mappable units of iron-formation also occur at two and possibly three different stratigraphic levels in the Mahnomen Formation and at two or more stratigraphic levels in the Rabbit Lake Formation. A depositional age for the North Range Group has not been established, but a Rb-Sr whole rock isochron age implies that it was metamorphosed at about 1,870 Ma (Peterman, 1966).

The Cuyuna South range structural terrane (loc. 5, fig. 17) just south of the Serpent Lake discontinuity consists mostly of mafic igneous rocks of both volcanic and hypabyssal affinity, carbonaceous slate, and lenses and layers of iron-formation (Morey and Morey, 1986). One of the volcanic units, the Randall Formation of Morey (1978), crops out at the southwest end of the terrane, together with thin-bedded, carbonate- and silicate-facies iron-formation.

The Animikie Group unconformably overlies the Mille Lacs and North Range Groups to the south and Archean rocks to the north. On the Mesabi (loc. 2, fig. 17) and Gunflint (loc. 1, fig. 17) ranges, it overlies Archean basement and consists of a basal quartz arenitic sequence, Kakabeka and Pokegama Quartzites; an intermediate iron-rich sequence, the Gunflint and Biwabik Iron-formations; and an upper thick graywacke-shale sequence, the Rove, Virginia, and Thomson Formations. Stratigraphic relationships are more complicated in the Emily district (loc. 3, fig. 17), where the stratigraphic position of the Biwabik is occupied by four lenticular units of iron-formation separated by intervening sequences of epiclastic rocks, either quartz arenite like the Pokegama Quartzite or black shale like the Virginia Formation (Morey and others, 1991). The principal iron-rich interval, unit A of the Ruth Lake area (Morey and others, 1991), contains an appreciable, but as yet unmeasured manganese resource.

The depositional age of the Animikie Group has not been established. However, Hemming and others (1990) have shown that the Pokegama Quartzite consists of Archean detritus that yields a Pb-Pb age of 2,612±3 Ma. This material is cut by quartz veins that are 1,930±25 Ma. The latter age is considerably younger than the Sm-Nd whole rock isochron age of 2,110±52 Ma reported by Gerlach and others (1988) for slaty units in the Biwabik Iron-formation. As judged from a Rb-Sr whole rock isochron age, the Thomson Formation along the south side of the Animikie basin was folded and metamorphosed at 1,770 Ma (Keighin and others, 1972).

A STRATIGRAPHIC SYNTHESIS

Rocks of the Penokean continental margin assemblage extend for more than 1,300 km along the south margin of the Superior craton. Sedimentation was diachronous over that distance: it began around 2,500 Ma in the Lake Huron region but did not begin until about 2,200 Ma in east-central Minnesota (fig. 19). Nonetheless, sedimentation had essentially ceased by ≈1,850 Ma in both areas. Thus, the Penokean continental margin assemblage apparently records 650 m.y. of geologic time, an interval greater than that attributed to all of the Phanerozoic Eon; the great length of time is a complicating factor for any stratigraphic synthesis.

Stratigraphic syntheses in Phanerozoic orogenic belts are typically based on biostratigraphic information that temporally constrains the preserved record of subsidence and sedimentation in marginal sedimentary prisms. However, biostratigraphic data are not everywhere available, even in the Phanerozoic, and the utilization of sequence stratigraphic concepts first developed by Sloss (1963) in recent years has greatly enhanced our understanding of relative-time stratigraphic relationships in these places. In particular Vail and others (1977) extended Sloss's (1963) concept of a sedimentary sequence to be applicable as the basic stratigraphic unit in shelf and coastal plain deposits. Sedimentary sequences consist of conformable and genetically related strata bounded at the top and bottom by unconformities or an equivalent laterally correlatable conformity. Thus unconformities assume considerable importance, not for the degree of erosion involved, but for their lateral persistence. It follows that where unconformity-bounded depositional sequences include isotopically datable rocks, the strata and their bounding surfaces may be calibrated in units of absolute time. Although this approach was developed in quite young Phanerozoic strata, it has been applied to Proterozoic strata with considerable success (Christie-Blick and others, 1988). The sequence-stratigraphic approach underlies the following discussion of the Penokean orogen.

The Huron Supergroup can be divided into two unconformity-bounded depositional sequences, which collectively were deposited during the interval between 2,491±5 Ma and 2,219±4 Ma. The chronostratigraphic significance of the lower Huronian–upper Huronian unconformity is uncertain, but Fairbairn and others (1969) have established a Rb-Sr isochron age of 2,240±87 Ma for the Gowganda Formation. Thus, the unconformity evolved before that time but after latest lower Huronian volcanism at 2,450±25 Ma.

Possible correlations between the Lake Huron and Lake Superior regions have been a nagging problem for more than 100 years. Young (1983) suggested that the Fern Creek and Enchantment Lake Formations in Michigan can be correlated with the Gowganda Formation in Ontario. If so, Early Proterozoic sedimentation may have started in the Lake Superior region as early as 2,327 Ma. Young (1983) also has correlated individual units of the Cobalt Group with individual units in the Chocolay Group. In that scheme the unconformity that tops the Chocolay Group is equivalent to the unconformity that tops the Cobalt Group, and all of the former would be older than 2,219±4 Ma. However, a second unconformity of unknown duration within the Chocolay Group separates the Fern Creek Formation from the remainder of the group (fig. 19). That unconformity also could be equivalent to the unconformity atop the Huron Supergroup. Therefore the bulk of the Chocolay Group could be either older or younger than 2,219±4 Ma.

Early Proterozoic rocks of glaciogenic affinity have not been recognized in Minnesota. However, the younger dolomitic rocks of the Trout Lake Formation and its associated unconformity at the top of the Mille Lacs Group in the Cuyuna North range are broadly similar to the Kona Dolomite and its associated unconformity at the top of the Chocolay Group in Wisconsin and possibly are correlative (Marsden, 1972). Building on Marsden's observation, Morey (1978) correlated the Mille Lacs and Chocolay Groups. The depositional age of volcanic strata in the lowermost part of the Mille Lacs Group in the Glen Township–Moose Lake structural terrane is well constrained at 2,197±39 Ma, an age that is statistically identical with the age of the Nipissing Diabase. Thus the unconformity beneath the Mille Lacs Group in the Glen Township–Moose Lake structural terrane likely is equivalent to the unconformity at the top of the Huron Supergroup. The Mille Lacs Group therefore is younger than the upper Huronian Cobalt Group. Because the bulk of the Chocolay Group has been correlated on lithic grounds with the Mille Lacs Group, it too must be younger than 2,219±4 Ma (fig. 19).

The North Range Group in east-central Minnesota and the Menominee Group in Michigan form two unconformity-bound stratigraphic sequences, which although different lithologically, are correlated here mainly because they occupy similar stratigraphic positions sandwiched between the Mille Lacs and Chocolay Groups below and the

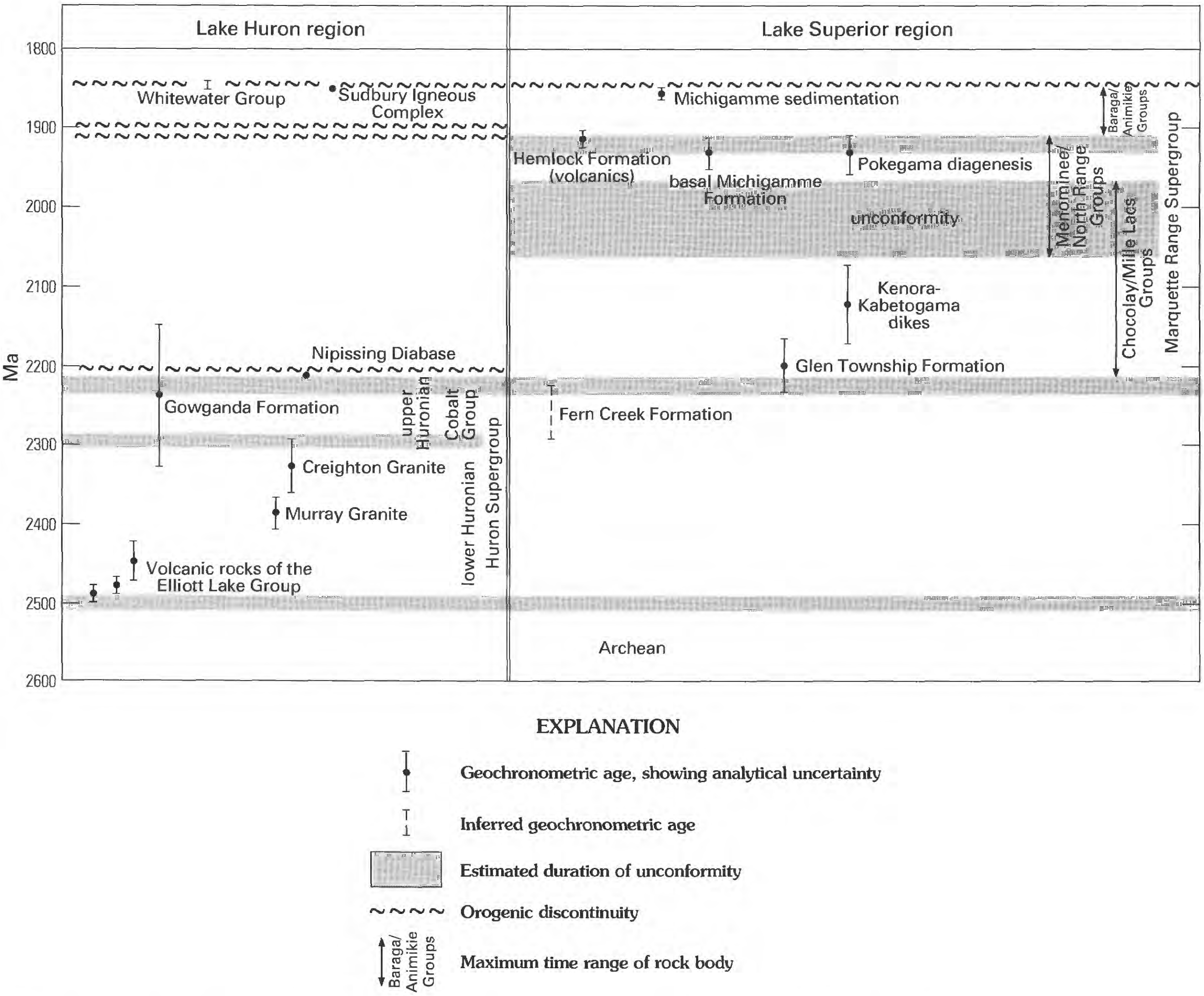


Figure 19. Geochronometric correlation chart of Early Proterozoic rocks in the Lake Huron and Lake Superior regions.

Animikie and Baraga Groups above. The amount of time represented by the lower bounding unconformity has not been established, but it could be as much as 150 m.y. However, volcanic rocks from the upper part of the Menominee Group yield an age of 1,910±10 Ma, indicating that sedimentation of it and the North Range Group was well underway by that time.

The unconformity at the top of the North Range Group is marked also by a structural break that separates twice-deformed rocks below from once-deformed rocks of the Animikie Group above. In contrast, the upper bounding surface of the Menominee Group is an unconformity in places but gradational in other places (Cannon, 1986), implying that the erosional surface, where developed, records a short interval of time. The short-lived nature of the Menominee-Baraga unconformity is further indicated by the fact that Baraga sedimentation was underway by 1,929±17 Ma, at about the time vein quartz was forming in the Animikie Group.

Broad stratigraphic similarities (for example, Morey, 1983b) between the Animikie Group in Minnesota and the Baraga Group on the western Gogebic range, as redefined herein, leave little doubt as to their correlation (fig. 20). Sedimentation culminated with the collision of arc-related rocks of the Wisconsin magmatic terranes with those of the continental margin along the Niagara fault zone. The precise timing of that event is uncertain, but a biotite-grade sample of a foredeep portion of the Michigamme Formation has yielded a U-Pb zircon age of 1,852±6 Ma (Sims and others, 1989).

Evidence of sedimentation in the Lake Huron region between the emplacement of the Nipissing Diabase at

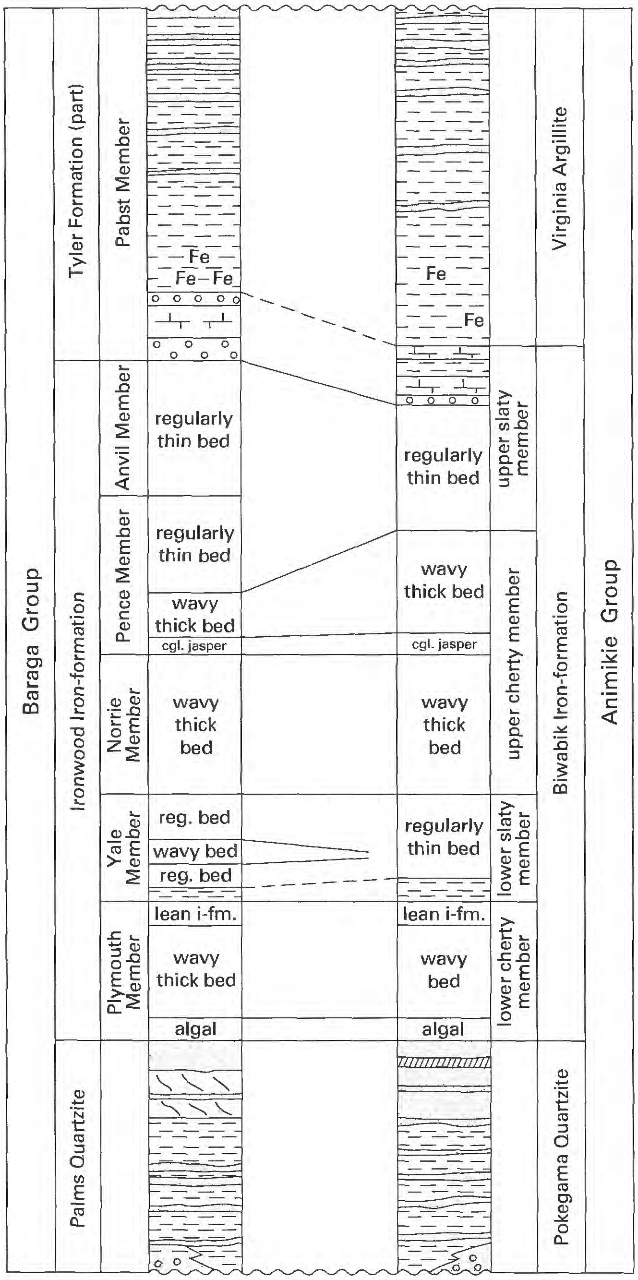


Figure 20. Inferred correlation of rock types in the Animikie Group on the Mesabi range in Minnesota with those of the Baraga Group, as redefined on the Gogebic range in Wisconsin; i-fm., iron-formation. Modified from Morey, 1983b; Ojakangas, 1983; Lucente and Morey, 1983; and Alwin, 1976.

2,210±4 Ma and the Sudbury Igneous Complex at 1,850±1 Ma is lacking. The Onaping and Chelmsford Formations of the Whitewater Group in the Sudbury basin have the attributes of a prograding flysch that Young (1983) suggested can be correlated with the upper parts of the Baraga and Animikie Groups in the Lake Superior region. Further, as in the Lake Superior region, sedimentation of the Whitewater Group was essentially over by about 1,850 Ma.

TECTONIC IMPLICATIONS

Numerous models have been proposed to explain the tectonic development of the Early Proterozoic sequences in the Lake Superior and Lake Huron regions. However, since about 1985 a consensus has existed that the Early Proterozoic rocks evolved within a plate tectonic framework wherein a continental margin formed, developed an early rift, received an assemblage of deposits, continued to grow and rift, and finally was destroyed (fig. 21).

The concept of a Penokean continental margin assemblage was developed by Schulz (1991) and Schulz and others (1993) for the Lake Superior region. They used geologic and geophysical observations from modern continental margins to establish a tectonic framework involving three stages—an intracratonic basin or intrarift stage, a rift stage, and a post-breakup stage. That framework was characterized by two major unconformities termed the "rift-onset unconformity" and the "break-up unconformity" of Falvey (1974). Thus, Schulz and others (1993) viewed the Marquette Range Supergroup and its correlatives as having been deposited as a continental margin assemblage in an extensional tectonic regime formed during the opening of a "Penokean Ocean," with the sequence segmented by a "rift-onset unconformity" atop the Mille Lacs and Chocolay Groups and a "break-up unconformity" atop the North Range and Menominee Groups. That model was modified subsequently by Southwick and others (1988) and Barovich and others (1989) to include the ideas of Hoffman (1987), who suggested that sedimentation of the Animikie and Baraga Groups took place in a northward-migrating foredeep basin formed in a compressional tectonic regime during closure of the "Penokean Ocean." The transition from continental margin sedimentation to foredeep sedimentation occurs somewhere within the thick turbidite deposits of the Animikie and Baraga Groups. I here extend that tectonic model to include the Lake Huron region.

Within the Penokean orogen, evidence of extensional tectonic processes is first recorded in the Lake Huron region, where lower Huronian fluvial and lacustrine sedimentary rocks were deposited in a series of fault-bounded basins (fig. 21*A*). Sedimentation was accompanied by bimodal volcanism and the emplacement of comagmatic hypabassal intrusions. Following a period of regional uplift and erosion, upper Huronian rocks were deposited over an angular unconformity of regional extent. Marine sedimentary patterns in the lower part of the upper Huronian were considerably modified by climate-dominated processes that produced various kinds of glaciogenic deposits (Young, 1991). The remainder of the upper Huronian rocks, however, record generally regressive sedimentary processes, probably related

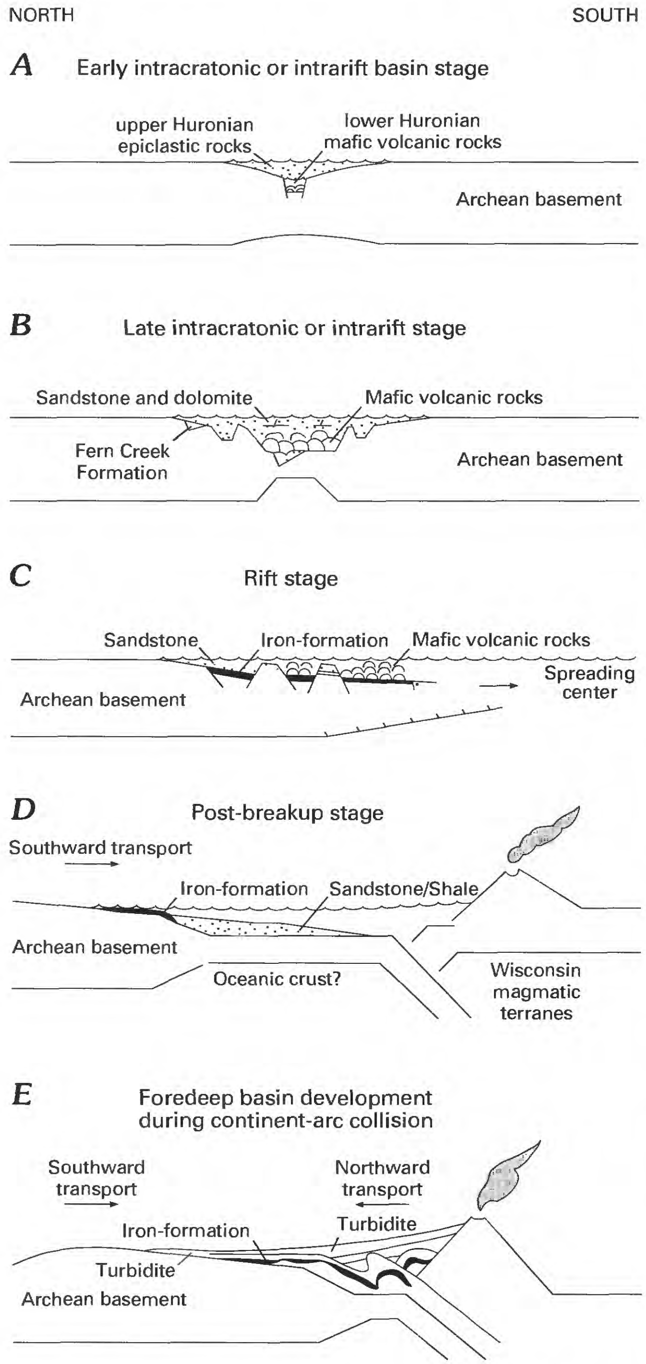


Figure 21. Schematic cross sections showing scenario for the tectonic evolution of continental margin assemblage rocks in the Penokean orogen. *A*, Huronian time (pre-2,220 Ma); *B*, Chocolay–Mille Lacs time (≈2,200 Ma); *C*, Menominee–Mille Lacs time; *D*, Early Baraga–Animikie time (≈1,900 Ma); *E*, Late Baraga–Animikie–Whitewater time (before 1,850 Ma). Modified from Barovich and others (1989).

to the gradual filling of the basin. Upper Huronian sedimentation was either terminated or closely followed by a compressional event that deformed the entire sedimentary package sometime before the Nipissing Diabase was emplaced at 2,219±4 Ma.

Glaciogenic deposits probably spread over parts of the Lake Superior region in pre-Nipissing time, but only small remnants are now preserved in structural lows, which are possibly fault-bounded depressions. However, much of the Chocolay Group was deposited under shallow-water conditions on a gradually subsiding platform or shelf much like that developed earlier in the Lake Huron region (fig. 21*B*).

Mille Lacs Group sedimentation in east-central Minnesota also occurred under fluvial to shallow marine conditions on a gradually subsiding shelf broken in places by fault-bounded basins where carbonate- and silicate-facies iron-formation accumulated with sulfide-rich black shale (fig. 21*B*). Sedimentation in these deeper water enclaves was accompanied by deposition of tholeiitic to calc-alkalic volcanic rocks that have geochemical signatures indicative of "within-plate" or "continental" tectonic settings (Southwick and Morey, 1991).

Sedimentary patterns change markedly across the "rift-onset unconformity" atop the Chocolay and Mille Lacs Groups. Deposition of the succeeding Menominee and North Range Groups marked the onset of a major period of rifting assigned to the so-called "rift stage" of modern evolving passive margins (fig. 21*C*). Consequently, sedimentation, including major units of iron-formation, was localized in structural basins bounded by normal listric faults (Cambray and others, 1991). Iron-formation in the Menominee Group accumulated together with tholeiitic basalt and lesser rhyolite that have geochemical signatures similar to Holocene "within-plate," rift-related "continental" basalts (Ueng and others, 1988; Beck and Murthy, 1991). Although thick units of eruptive rocks did not accumulate in the North Range Group, the Trommald Formation does contain evidence of extensive hydrothermal fumarolic activity (Morey and others, 1992a,b; McSwiggen and others, 1992), including beds of tourmalinite (Boerboom, 1989). Clearly the North Range Group, like the Menominee Group, accumulated in a fault-bounded basin formed in response to regional extensional stresses.

The ultimate initiation of sea-floor spreading terminated active rift development and initiated deposition of the lower parts of the Baraga and Animikie Groups on a surface of regional extent (fig. 21*D*) termed the "break-up unconformity." The transition from the "rift stage" to the "post-breakup stage" in the Lake Superior region must have occurred fairly quickly, because isotopic ages from the upper part of the Menominee Group and the lower parts of the Baraga and Animikie Groups cluster around 1,900 Ma.

The transition from extensional to compressional stresses and the onset of foredeep deposition occurred during deposition of the Animikie and Baraga Groups (fig. 21*E*). The lower part of the Animikie Group contains detritus derived from an Archean craton (Hemming and others,

1990) to the north, which was deposited, along with iron-formation and a succeeding black shale facies, on a rapidly subsiding continental shelf (Ojakangas, 1983). The lower part of the Baraga Group also was derived from an Archean crustal source as evidenced by the neodymium isotopic studies of Barovich and others (1989). In contrast, graywacke from the upper part of the Baraga Group was derived from an Early Proterozoic source that Barovich and others (1989) suggested was high ground formed as island arc-related rocks of the Pembine-Wausau terrane of the Wisconsin magmatic terranes were accreted to the continental margin.

A significant unconformity or other stratigraphic markers have not been identified that would define the transition from post-breakup to foredeep basin sedimentation in the Baraga or Animikie Groups. Nonetheless, some chronostratigraphic limits can be placed on the Baraga Group itself. Sedimentation started at approximately 1,929±17 Ma and was over by 1,852±6 Ma, as judged from a U-Pb zircon age obtained from biotite-grade Michigamme Formation known from Nd isotopic data to have an Early Proterozoic source. Chronostratigraphic limits on the Animikie Group are less well constrained, but deposition must have occurred between 1,930±25 Ma and approximately 1,770 Ma, the time of a mild metamorphic event that affected the southern part of the Animikie Group in east-central Minnesota. However, other lines of evidence imply that compressional tectonic processes were well underway in east-central Minnesota by 1,870 Ma, when strata of the North Range Group were metamorphosed and several syntectonic intrusions were emplaced in the McGrath–Little Falls structural terrane.

Geochronometric evidence from the Lake Huron region suggests that the Whitewater Group accumulated over no more than 1 or 2 m.y. of geologic time about 1,850 Ma, when sedimentation was abruptly terminated by Penokean folding and metamorphism. If the upper parts of the Baraga/Animikie and Whitewater Groups are correlative, as suggested by Young (1983), the combined geochronometric evidence implies that foredeep sedimentation had ended in the Lake Superior region by 1,852 Ma, when the basin fill was consumed along the trailing edge of the northward-migrating foredeep. At the same time, however, the leading edge of that foredeep basin was still active in the Lake Huron region and was still receiving detritus from an Archean craton to the north. Sedimentation of the Whitewater Group ended shortly after 1,850 Ma, marking the end of the Penokean orogeny in the Lake Huron region.

STRUCTURE OF CONTINENTAL MARGIN

By P.K. Sims

Geologic mapping and structural analysis during the past decade have established that the Early Proterozoic continental margin in the Lake Superior region is a foreland fold-and-thrust belt (Southwick and Morey, 1991; Klasner and Sims, 1993) analogous to younger foreland belts such as the northern Appalachians (Gregg, 1993). Collision along the Niagara (suture) fault zone in northern Michigan and Wisconsin and its presumed equivalent in east-central Minnesota has produced predominantly north-verging folds that have been transported northward along imbricate thrusts. In northern Michigan, the Archean basement was extensively involved in the thrusting. In general aspects, the continental margin of the Penokean orogen consists of a parautochthonous, little-deformed terrane on the north, adjacent to and overlying exposed Archean rocks of the Superior Archean craton—the cratonic foreland—and a southern fold-and-thrust belt, adjacent to the Niagara fault zone.

NORTHERN MICHIGAN AND ADJACENT WISCONSIN

In northern Michigan and adjacent Wisconsin, the terrane assemblage on the Early Proterozoic continental margin, from north to south, consists of a parautochthonous and allochthonous foreland (Gregg, 1993), mainly involving thin-skinned deformation, a complex uplifted zone of Archean rocks, including gneiss domes, a zone of out-of-sequence backthrusting involving thick-skinned deformation (Klasner and Sims, 1993), followed by a zone of in-sequence, northward-directed thrusting, which includes the Niagara (suture) fault zone itself.

In the parautochthonous terrane within the Baraga belt, north of the Falls River thrust (fig. 22), slaty rocks of the Michigamme Formation (Xm_2) contain a single fold set (F_1) and associated cleavage (Gregg, 1993). Small-scale, synkinematic north-verging thrusts deform the folds and cleavage. The thrusts are parallel to and apparently related to the Falls River thrust, which strikes approximately east and dips 30°–40° S. and which separates the parautochthonous rocks from allochthonous rocks to the south.

The allochthonous rocks to the south of the Falls River thrust, referred to as the Falls River slice (Gregg, 1993), are slaty rocks of the Michigamme Formation that contain north-verging thrusts and associated folds (F_1) overprinted by D_2 folds (F_2) and S_2 crenulation cleavage. Field relations indicate a late D_1 timing for the thrusting in both the terranes on opposite sides of the Falls River thrust. During D_2 deformation, early first deformation structures were rotated to nearly recumbent positions. The finite strain associated with D_1 is greater in the Falls River slice than in the parautochthon to the north (Gregg, 1993). Gregg has suggested that the Falls River thrust correlates with a similar boundary recognized by Holst (1984) in east-central Minnesota, referred to as a "nappe front," and discussed following.

The southern boundary of the Falls River slice can be taken as the Covington thrust, recognized by Gregg (1993) in an exposed section of the Michigamme Formation along U.S. Highway 141 about 4.5 km south of Covington (fig. 22). The south-dipping fault is marked by a prominent bedding-cleavage intersection lineation that plunges steeply south; the cleavage is oriented differently on opposite sides of the fault. The fault coincides with the boundary between two structural domains delineated by Foose (1981) in the northwest corner of the Ned Lake 15-minute quadrangle, a few kilometers to the east of U.S. Highway 141. In the block to the north, folds plunge gently east or west and axial surfaces dip steeply south; in the block to the south, folds also plunge gently east or west but axial surfaces dip steeply north. That the Covington thrust is probably a major structure is indicated by Nd isotopic data. Samples of Michigamme Formation to the north of the fault and samples of pre-Michigamme Formation rocks in the Marquette syncline, to the east, indicate an Archean source for the sediment, presumably from the craton to the north, whereas samples of the Michigamme Formation to the south of the fault indicate an Early Proterozoic source and deposition in foredeep basins (Barovich and others, 1989). The Covington thrust presumably marks the northern limit in Upper Michigan of imbricate thrusts on which foredeep deposits have been transported northward over craton-margin deposits.

Southeast of the Covington thrust, the Archean basement is definitely involved in the fold-and-thrust deformation (Klasner and Sims, 1993). The extent to which the Archean rocks in the Amasa uplift and the southern complex (fig. 22) are allochthonous, however, is uncertain. Gregg (1993) has inferred that indeed both the Amasa uplift and the southern complex of the Marquette district (gneiss domes) are allochthonous, as is the northern complex, but this interpretation has not been verified by detailed structural studies.

South of the Amasa and southern complex domes, the Bush Lake fault defines the northern limit of out-of-sequence backthrusting (Klasner and Sims, 1993). The Bush Lake fault is a major north-verging, high-angle thrust that brings Archean basement and Early Proterozoic rocks (Marquette Range Supergroup) partially over the Amasa dome (fig. 22). Figure 23 is a diagrammatic sketch showing the inferred structural relationships of backthrusting. Associated with the backthrusting, and presumably related to it, is the Carney Lake nappe (Sims and Schulz, 1993), the upper limb of which apparently has been eroded.

The cause of the backthrusting is equivocal. Morley (1988) has suggested that backthrusts as out-of-sequence faults can form in foreland thrust belts at an abrupt change in thickness of the thrust sheet, that is, they form where the critical taper of the thrust wedge decreases and forward propagation of the thrust sheet cannot be maintained. Hence, the thrust wedge is thickened by backthrusting and internal folding until a critical taper is accomplished and the in-sequence forward thrusting can continue. Alternatively, the foreland propagation of a thrust may be stopped by a buttress, such as the Amasa uplift. Such backthrusts are not uncommon; they have been described from the Canadian and U.S. Rockies (Price, 1986; Tysdal, 1986), the Alps (Tricart, 1984), and elsewhere.

The geometric pattern of thrust faults on the southernmost part of the continental foreland (fig. 22) suggests that the episode of backthrusting was followed temporally by north- to northeast-directed thrusting that impinged laterally on the previously folded and thrust rocks. This scenario suggests that collision along the Niagara fault zone was diachronous, initially being directed northward, followed by northeastward vergence in the southernmost area of figure 22. This conclusion is supported by two types of field observations. On the west side of the Carney Lake nappe, Higgins (1947) mapped several steep, east-trending transcurrent faults that offset the steeply dipping Sturgeon Quartzite (unit Xm_1, fig. 22) and which postdate the principal folding, now recognized as nappe-forming deformation. These transcurrent faults have gently west plunging slickenlines, suggesting upward and eastward movement on the faults. They can be attributed to northeast-directed compression resulting from impingement of the Badwater thrust fault slice on the Carney Lake block. Secondly, in the Amasa uplift, northeast-verging faults, interpreted mainly as thrusts, and related small-scale folds (east of Crystal Falls) are transverse to the older northwest-trending structural grain of the Amasa uplift (Bayley, 1959; Sims and Schulz, 1993), and clearly are younger than the major deformation. Thrusting could have been on a scale sufficient to imbricate the Hemlock Formation (unit Xm_1, fig. 22) on the west side of the dome, greatly increasing its thickness (Foose, 1981), and also to cause reorientation of the Archean core and flanking younger rock from an original east-west trend to a northwest trend.

EAST-CENTRAL MINNESOTA

Southwick and others (1988) and Southwick and Morey (1991), following earlier studies of Holst (1982, 1984) and Holm and others (1988), have documented the structure in the poorly exposed continental foreland on the west side of the Midcontinent rift system, based mainly on detailed aeromagnetic data, gravity data, and drilling, together with sparse outcrop data. They concluded that the Penokean orogen in this area consists of an internal fold-and-thrust belt

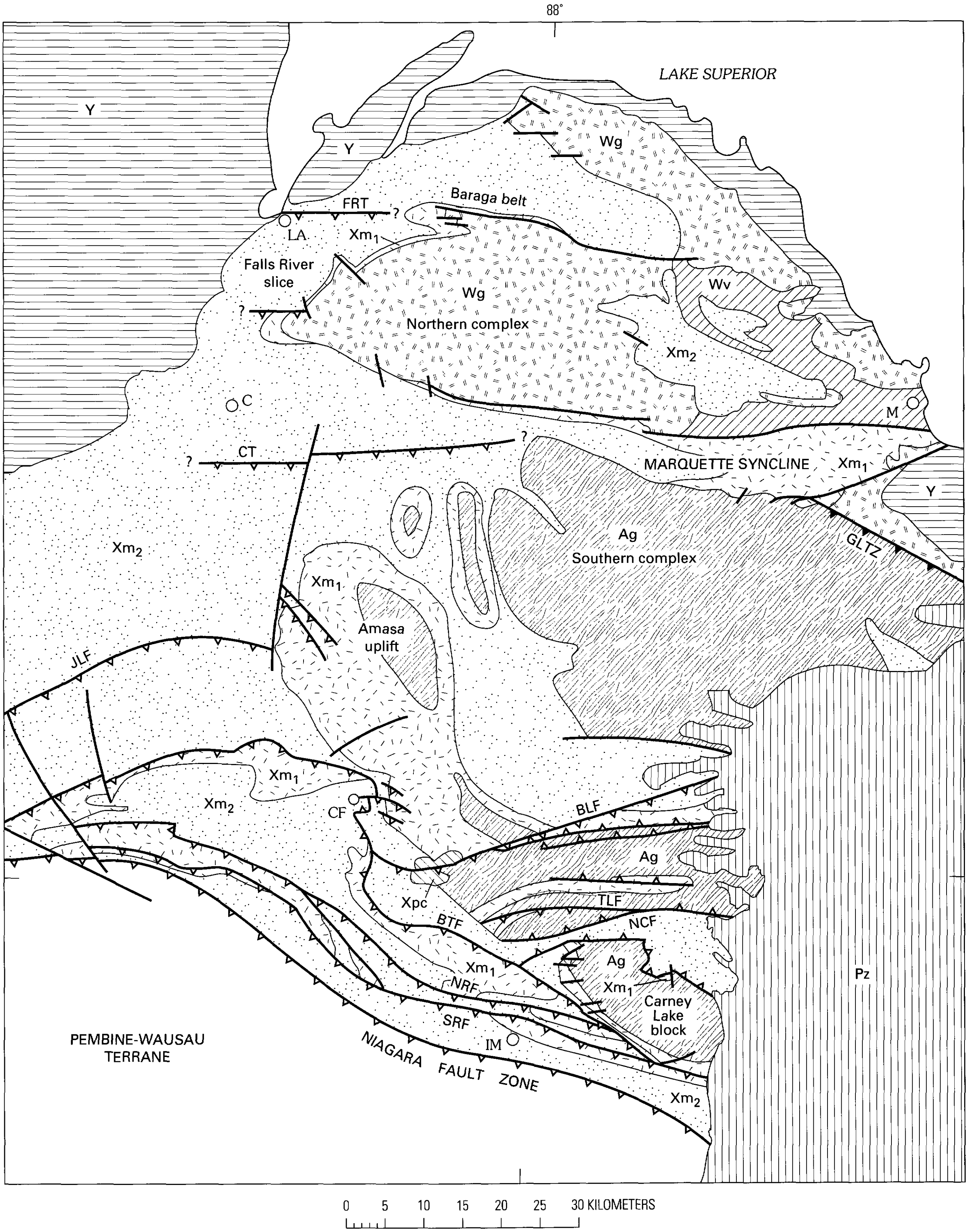


Figure 22 (above and facing page). Generalized structure map of continental margin of Penokean orogen in Upper Michigan. C, Covington; CF, Crystal Falls; IM, Iron Mountain, LA, L'Anse; M, Marquette. Compiled by P.K. Sims, 1991.

EXPLANATION

Pz — Paleozoic rocks

Y — Rocks of Midcontinent rift system (Middle Proterozoic ≈1.1 Ga)

EARLY PROTEROZOIC (1.6–2.5 Ga)

Xpc — Peavy Pond Complex

Marquette Range Supergroup

Xm_2 — Michigamme Formation and correlative rocks; locally includes unit Xm_1

Xm_1 — Pre-Michigamme rocks

ARCHEAN (2.5 Ga and older)

Greenstone-granite terrane (Wawa subprovince)

Wg — Granitoid rocks (2.65–2.75 Ga)

Wv — Metavolcanic and metasedimentary rocks

Gneiss terrane (2.6–3.6 Ga) (Minnesota River Valley subprovince)

Ag — Gneiss, amphibolite, and migmatite

Contact

High-angle fault

Thrust fault (Early Proterozoic)—Queried where extent uncertain; sawteeth on upper plate

Thrust fault (Late Archean)—Sawteeth on upper plate

GLTZ — Great Lakes tectonic zone

BLF — Bush Lake fault
BTF — Badwater thrust fault
CT — Covington thrust
FRT — Falls River thrust
JLF — James Lake fault
NCF — North Calumet fault
NRF — North Range fault
SRF — South Range fault
TLF — Toby Lake fault

(on the south) that contains tectonically imbricated sedimentary and volcanic rocks in several terranes of contrasting stratigraphy, structural style, and metamorphic grade, together with an external turbidite basin (Animikie basin) that they interpreted as a migrating foredeep (fig. 24). A master fault, named the Malmo discontinuity, separates definite continental margin rocks from an enigmatic, southern (McGrath–Little Falls) terrane of syntectonic and post-tectonic granitoid rocks that intrude metamorphic rocks of both Archean and Early Proterozoic ages. They suggested that the Malmo discontinuity correlates with the Niagara fault zone, on the east side of the Midcontinent rift system. The tectonic zonation of these authors corresponds roughly with that distinguished in the better exposed area in Upper Michigan, but substantial differences exist.

Sedimentary strata of the Animikie Group, which taken together compose the external turbidite basin of Southwick and others (1988), lie unconformably on the Archean craton on the northwest side of the Animikie basin, and with inferred unconformity on previously deformed rocks of the fold-and-thrust belt on the southeast side (fig. 24). Strata near the southern flank of the basin were folded when the underlying rocks of the fold-and-thrust belt were refolded during the later stages of regional compression, whereas they were scarcely deformed on the cratonal northern flank. The boundary between the little-deformed and multiply deformed rocks, near Carlton, corresponds approximately with the "nappe front" of Holst (1984). The rocks on the north side of the Animikie basin have a northerly (cratonal) source, whereas those near the southern basin margin are consistent with a southern provenance (Southwick and others, 1988). The broad features of the Animikie basin and smaller outliers have been interpreted by Southwick and colleagues to have formed from a migrating foredeep that overrode and was at least partly deposited on older rocks of the fold-and-thrust belt.

The fold-and-thrust belt of the Penokean orogen in Minnesota is made up of four major structural terranes (fig. 24), which are separated by major discontinuities, presumably north-verging thrust faults. The southernmost terrane, the McGrath–Little Falls terrane, has been interpreted by Southwick and Morey (1991) as composed of volcanic arc, perhaps accreted arcs, and continental fragments; thus, in a strict sense, it is not a part of the continental margin assemblage. Instead, it is more nearly analogous to the Wisconsin magmatic terranes. The McGrath–Little Falls terrane comprises relatively high grade gneiss and schist, a mantled gneiss dome (McGrath dome), and several late-tectonic to post-tectonic granitoid plutons. To the north of the McGrath–Little Falls terrane and separated from it by the Malmo discontinuity are the Glen Township–Moose Lake terrane and adjacent Cuyuna South range terrane. Both of these terranes contain folded volcanic and sedimentary rocks, mainly of the Mille Lacs Group. The rocks are multiply deformed, as determined previously by Holst (1982, 1984). The geochemistry of the volcanic rocks of the Mille Lacs Group supports an extensional setting in a continental environment (Southwick and Morey, 1991). Northwest of the Glen Township–Moose Lake and Cuyuna South range terranes, the Cuyuna North range terrane contains weakly metamorphosed, less strongly deformed sedimentary rocks that are interpreted as having formed in a small restricted basin that was incorporated tectonically in the more external part of the fold-and-thrust belt. As noted previously, the

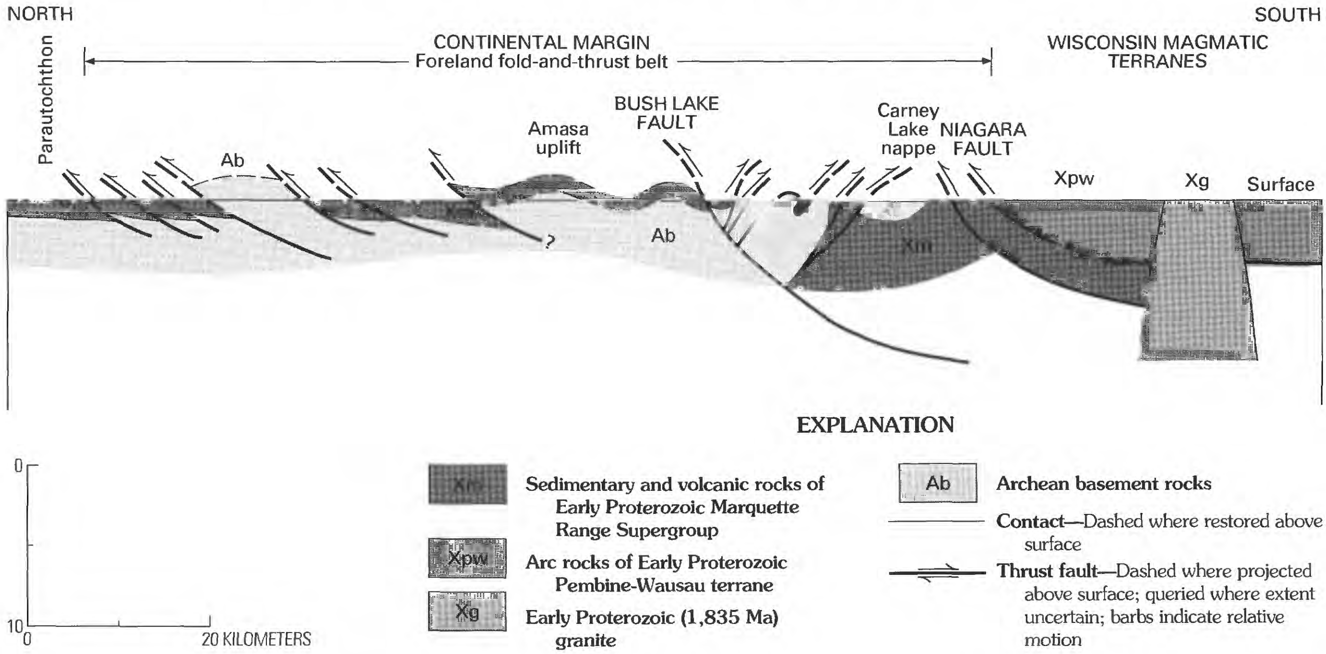


Figure 23. Diagrammatic structure section across Early Proterozoic continental margin, northern Michigan. Modified from Klasner and Sims (1993). North-south section through Amasa uplift.

fold-and-thrust belt abuts, and is partially overlapped by, the turbidite-filled Animikie basin.

COMPARATIVE STRUCTURAL GEOLOGY

Although generally comparable in the large view, structures in the rocks of the continental margin assemblage on opposite sides of the Midcontinent rift system differ in some respects. In northern Michigan, parautochthonous and definite allochthonous rocks are separated by a north-verging thrust fault, the Falls River thrust (fig. 22; Gregg, 1993). The parautochthonous rocks as well as adjacent strata of the lower part of the Marquette Range Supergroup in the Falls River slice are continental margin sedimentary rocks of a northern, Archean provenance (Barovich and others, 1989). The younger, foredeep deposits, as indicated by Nd isotopic data, are apparently juxtaposed with the passive-margin sedimentary rocks, to the north, along the Covington thrust, indicating that these foredeep strata have been transported northward along imbricate thrust faults.

In central Minnesota, on the other hand, the little-deformed foredeep deposits, such as in the Animikie basin, overlie craton-derived sediments in the north margin of the basin. Existence of foredeep deposits in the fold-and-thrust belt has not been confirmed, but they should be present at least locally.

The differences in Penokean structure and foredeep deposits on opposite sides of the Middle Proterozoic rift system are to be expected because of the dynamic conditions under which foredeeps form. The turbiditic sedimentary rocks within the basins should provide a record of changing provenance, current regimes, and slope conditions as a function of time, especially near their south margins. Cryptic unconformities between physically similar turbidite wedges should exist in the Penokean foredeeps, but none has yet been demonstrated. The probability of their existence should guide future stratigraphic studies in the Animikie and Michigan basins and be factored into tectonostratigraphic reconstructions.

The suggested equivalence (Southwick and others, 1988) of the Niagara fault zone and the Malmo discontinuity needs further testing. The Niagara fault zone is well established as a tectonic suture between continental margin rocks on the north and rocks of island-arc affinity (Wisconsin magmatic terranes) on the south (Sims and others, 1989, and references therein). Archean basement rocks are not known in the northern Pembine-Wausau arc terrane, but are widespread in the southern Marshfield terrane, where they at least partly underlie Early Proterozoic calc-alkaline volcanic and granitoid rocks. The apparent westward convergence of the Niagara fault zone and the Eau Pleine shear zone in northern Wisconsin and the apparent wedging out of the Pembine-Wausau terrane westward (pl. 1) could account for the apparent absence of arc-related volcanic rocks in the age range 1,890–1,840 Ma in the magmatic (McGrath–Little Falls) terrane (unit XAi, pl. 1) in Minnesota. The existence of comparable calc-alkaline intrusive rocks in both structural terranes south of the Niagara and Malmo structures, and

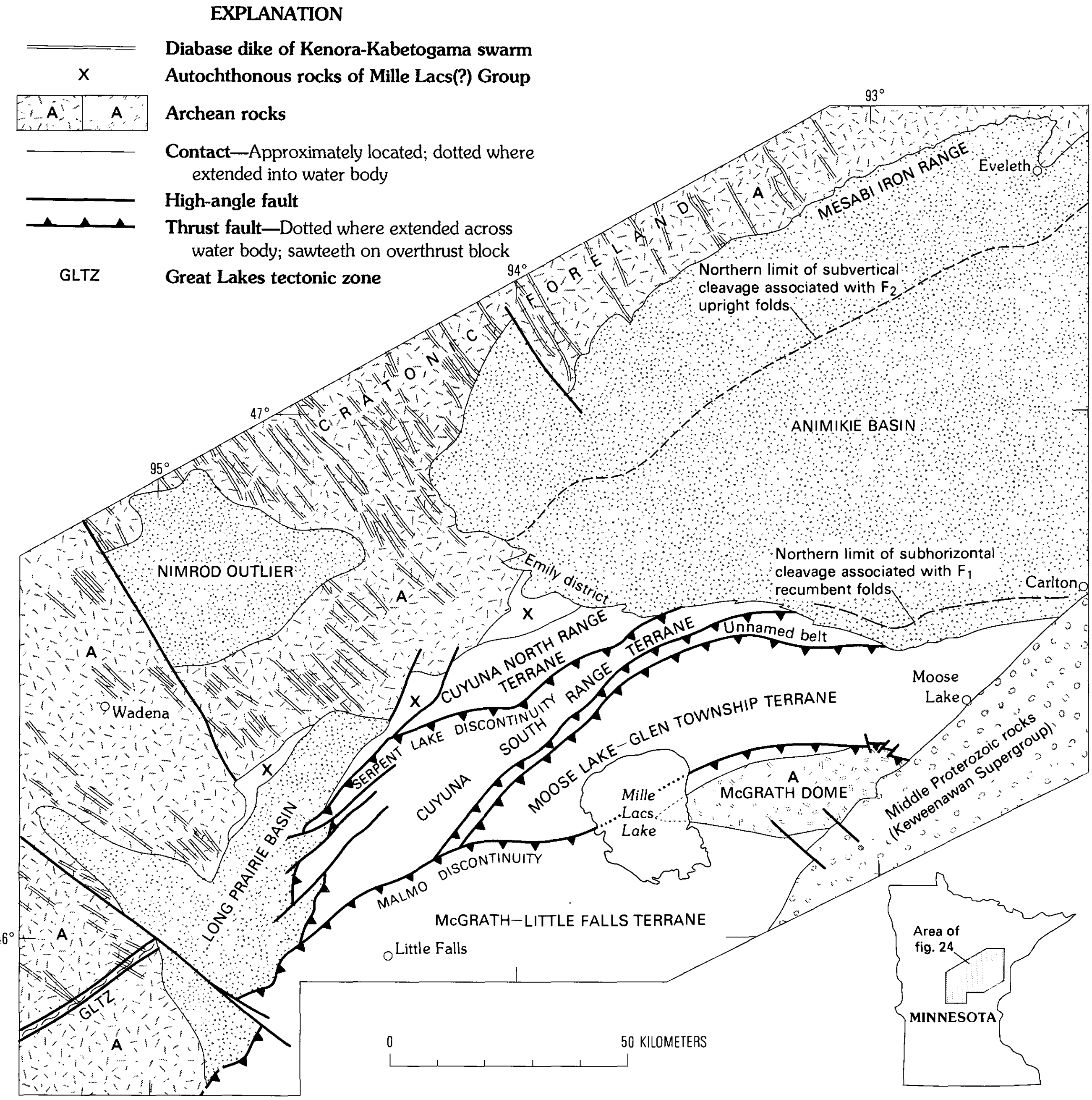


Figure 24. Generalized tectonic map of Penokean orogen in east-central Minnesota, showing major tectonic subdivisions. Long-dashed line demarks northern limit of recumbent deformation style (F_1) as mapped by Holst (1982, 1984) and extended by Southwick and Morey (1991). Blank areas, mainly Early Proterozoic rocks of fold-and-thrust belt. Modified from Southwick and Morey (1991).

their near absence to the north, favors a correlation of these terranes across the rift system. If this correlation is valid, however, the Archean gneiss in the McGrath dome and in areas farther south (Southwick and others, 1988) must be allochthonous, and therefore not correlative with the gneisses in southwestern Minnesota that compose the vast Archean gneiss terrane (Minnesota River Valley subprovince of Superior province). That the gneiss in the McGrath dome is indeed probably allochthonous is indicated by magnetotelluric studies by Wunderman and Young (1987). An anomaly associated with the Malmo zone in Minnesota dips southward beneath the McGrath dome. This magnetotelluric anomaly appears to correlate with the Flambeau anomaly in northwestern Wisconsin (Sternberg and Clay, 1977), which strengthens the suggested correlation of the Malmo and Niagara zones.

AGE CONSTRAINTS ON PENOKEAN DEFORMATION

The beginning of Penokean deformation is poorly constrained in the Lake Superior region, but the termination of deformation is accurately dated. Deformation could have started any time after 2,197±39 Ma (Beck, 1988), the time of eruption of continental margin volcanic rocks of the Mille Lacs Group in east-central Minnesota. In north-central Minnesota, it is known to be younger than 2,125±45 Ma, the age of a northwest-trending diabase dike (Kenora-Kabetogama) swarm in Archean basement rocks that is unconformably overlain by basal units of the Animikie Group (Southwick and Halls, 1987). In northern Michigan, deformation started sometime after 1,910±10 Ma (Van Schmus and Bickford, 1981), the age of a sample of the Hemlock Formation of the Menominee Group.

In Michigan, age constraints on the timing of collision along the Niagara fault zone are provided by a U-Pb zircon age of 1,852±6 Ma (Sims, Peterman, and Schulz, 1985) on a sample of biotite schist (Michigamme Formation) known from Nd isotopic data to have had an Early Proterozoic source, presumably eroded rocks of the Pembine-Wausau terrane. The time of graywacke deposition dates the final convergence of the Pembine-Wausau terrane with the continental margin, as well as the culmination of foredeep development. This age is about 10 m.y. younger than that of foliated granitoid rocks in the Dunbar area, within the Pembine-Wausau terrane (Sims and others, 1992). The age of post-tectonic stitching plutons (Bush Lake and Spikehorn Creek Granites), 1,835±6 Ma (Sims, Peterman, and Schulz, 1985), inferred to pierce the Niagara fault zone in northeastern Wisconsin (Sims and Schulz, 1993; Xg, fig. 23), indicates that collision at this locality definitely predated 1,835 Ma. Similarly, in Michigan, rare, small post-tectonic granitic plutons have an approximate age of 1,824 Ma.

In Minnesota, deformation was underway at 1,869±5 Ma (Goldich and Fischer, 1986), the age of a syntectonic pluton (Bradbury Creek Granodiorite of Morey, 1978)

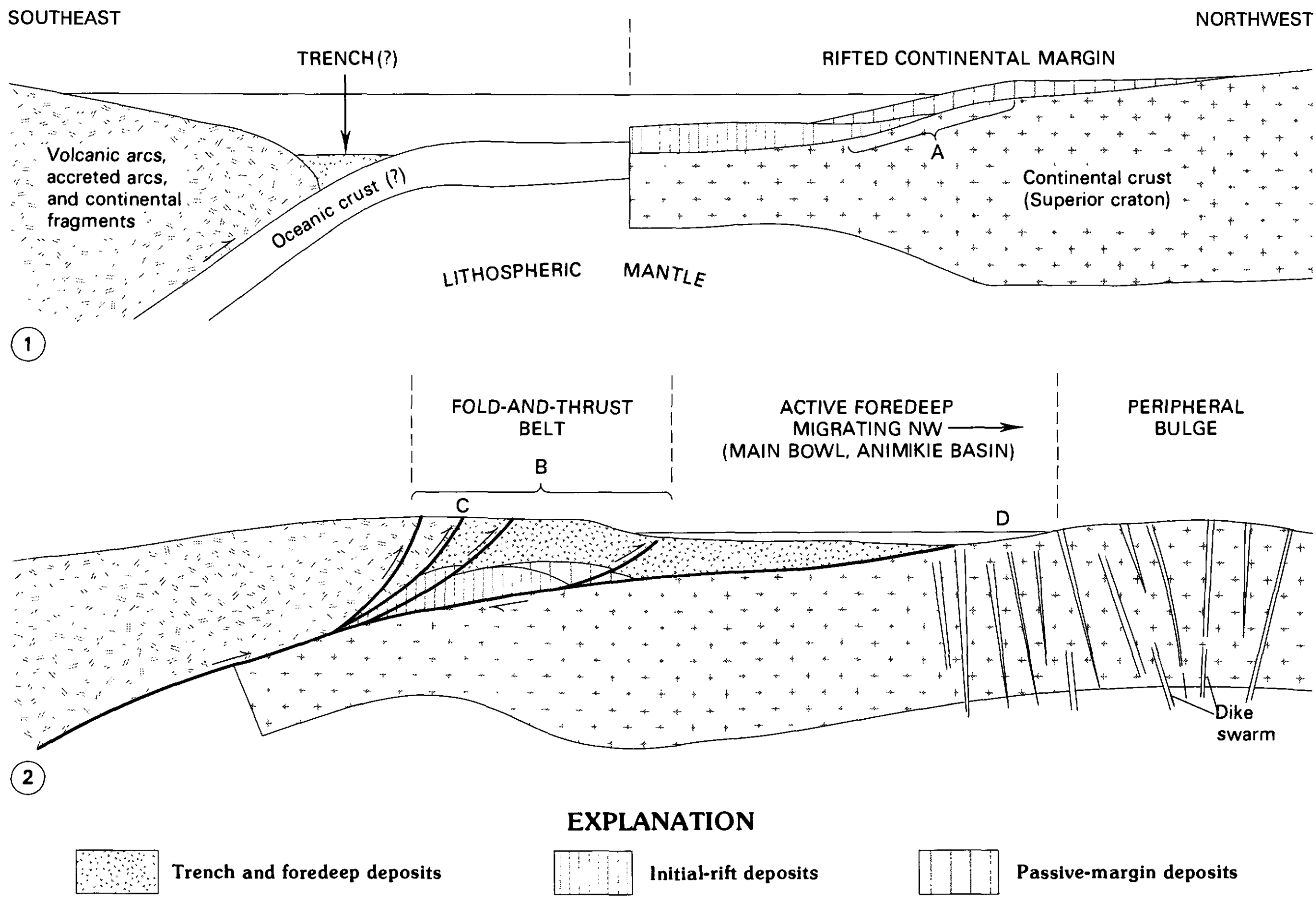


Figure 25. Schematic diagram modified from Hoffman (1987) illustrating the evolution of an oceanic trench (stage 1) into a foredeep (stage 2). A, passive-margin deposits accumulating on continental edge; B, location of fold-and-thrust belt involving volcanic and sedimentary rocks and their basement; C, Niagara fault zone; D, passive-margin deposits. Fold-and-thrust belt is approximately 50 km wide. Modified from version in Southwick and Morey (1991).

emplaced into complex migmatite (Hillman Migmatite of Morey, 1978), and had ceased by 1,812±9 Ma (S.S. Goldich, cited in Horan and others, 1987), when the post-tectonic Rockville and Reformatory Granites of Morey (1978) were intruded, mainly in the McGrath–Little Falls structural terrane (fig. 24). St. Cloud Granite plutons (≈1,770 Ma; Spencer and Hanson, 1984) also occur in the southernmost terrane, and have been interpreted to pierce and "pin" the imbricate stack.

Age constraints on gneiss doming in Upper Michigan are provided by secondary whole-rock and mineral Rb-Sr isochron ages in the range 1,800–1,750 Ma on Archean and Early Proterozoic rocks from the Watersmeet dome (Sims and others, 1984). The gneiss doming reflects the rise of isotherms consequent on the stacking of imbricate thrust sheets on the continental margin (Attoh and Klasner, 1989). The resetting of ages in the domes occurred 50 to 75 m.y. after suturing along the Niagara and Malmo structural zones.

From evidence in the Wisconsin magmatic terranes, Sims and others (1989) have concluded that Penokean deformation in this area had ceased before 1,835 Ma, when post-tectonic red alkali-feldspar granite intruded the Eau Pleine shear zone (fig. 26, unit Xgr), the suture between the Pembine-Wausau and Marshfield terranes.

TECTONIC MODEL

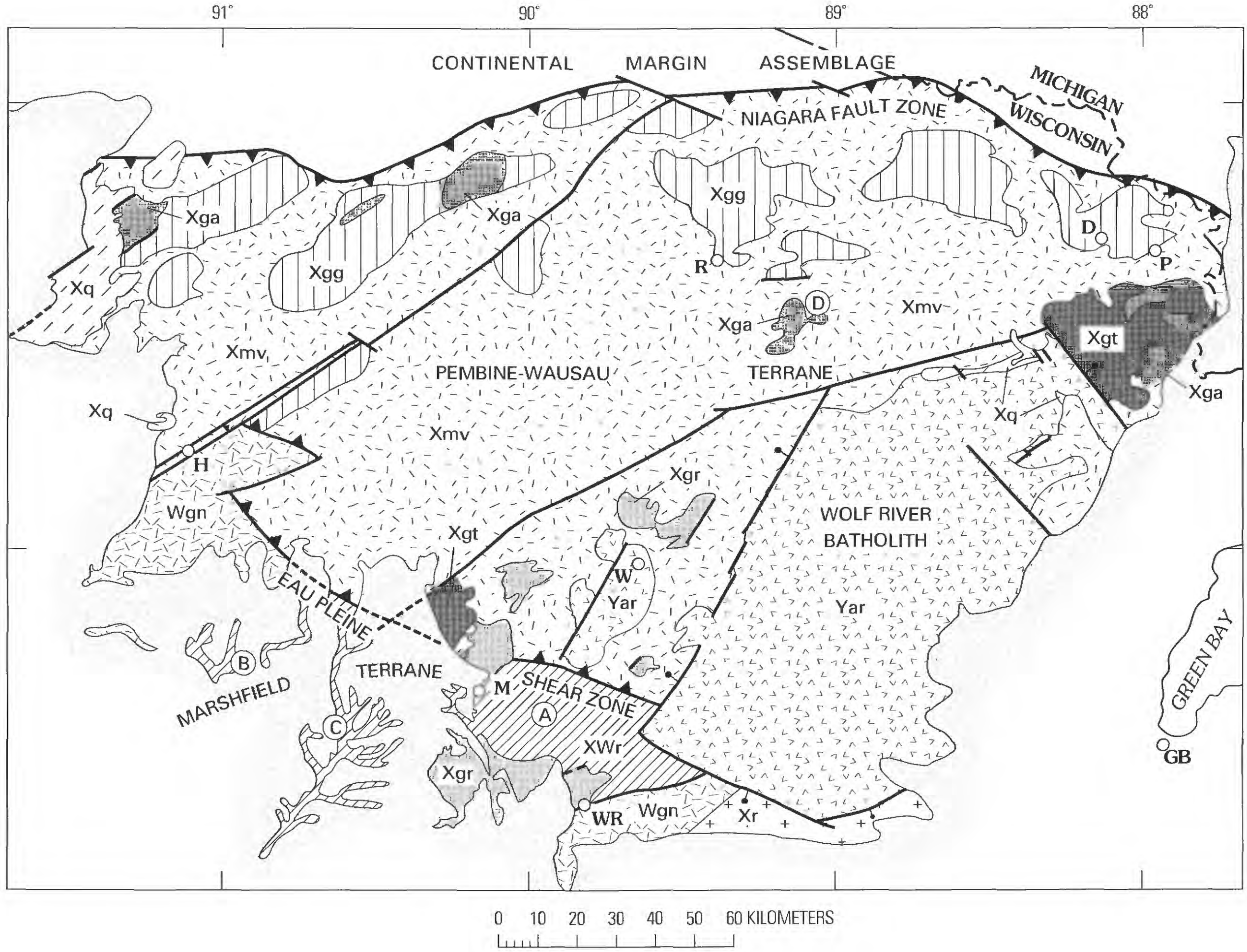
The conceptual tectonic model proposed by Southwick and Morey (1991) for Minnesota is applicable, with slight modification, to the Penokean orogen as a whole in the Lake Superior region (fig. 25). Stage 1 of figure 25 represents the rifted margin of the Archean Superior craton approaching a south-dipping subduction zone in Early Proterozoic time. Initial-rift deposits, including bimodal volcanic rocks, are riding on the continental edge; passive-margin deposits, including quartzite and dolomite of the Chocolay Group and iron-formation and quartzite of the Menominee Group in Michigan, and quartzite in the Mille Lacs Group in Minnesota, are accumulating near A. In both areas, and especially in Michigan, the passive-margin deposits include a slate-graywacke succession that is preserved along and near the north margins of the Animikie and Michigan basins. Stage 2 of figure 25 represents active subduction of the continental margin. Crustal loading by the ensuing fold-thrust mass has occurred, generating as a flexural response an actively migrating foredeep (including Animikie basin and, in Michigan, the Baraga belt (fig. 22)) on the continental foreland. The initial-rift and passive-margin deposits and, in Michigan at least, the Archean basement, as well as the foredeep deposits are involved in the fold-and-thrust deformation (B, fig. 25). "Fold nappes" (Ramsay, 1967), such as the Carney Lake nappe in Michigan (Sims and Schulz, 1993) and the recumbent nappe described by Holst (1984) in Minnesota, developed outboard on the continental foreland, but "thrust nappes" (Ramsay and Huber, 1987) were the major structures developed. In their model for the Minnesota segment, Southwick and Morey (1991) included a slice of the southern arc (McGrath–Little Falls) terrane in the fold-and-thrust deformation, whereas the suture on the east side of the rift juxtaposes arc and continental terranes. Locality C in figure 25 diagrammatically represents the Niagara fault zone in this area. Another modification in the model required for Michigan is the preservation of passive-margin (graywacke-shale) deposits on the north (cratonic) side of the Baraga belt (D, fig. 25).

An additional element of the model for the eastern segment of the Penokean orogen was the development of a north-dipping subduction zone (Eau Pleine shear zone) about 15 m.y. after final collision along the Niagara suture. This suture marked the closing of an ocean basin between the accreted arc (Pembine-Wausau) terrane and a minicontinent (Marshfield terrane) on the south (pl. 1). This suturing postdated the Penokean orogeny.

WISCONSIN MAGMATIC TERRANES

By P.K. Sims *and* K.J. Schulz

The Wisconsin magmatic terranes are complex volcanoplutonic sequences that compose two terranes (Sims and others, 1989). A northern terrane, named the Pembine-Wausau (P-W) terrane, is separated from a southern (Marshfield) terrane by the Eau Pleine shear zone, a south-verging paleosuture (Sims, 1989; 1990a). After amalgamation of the two magmatic terranes at about 1,840 Ma, post-tectonic alkali-feldspar granite (1,835 Ma) was intruded as stitching plutons, and cogenetic silicic rhyolite was erupted from possible calderas in the vicinity of the Eau Pleine shear zone (Sims, 1990a). Later, small 1,760 Ma granitoid plutons were emplaced in the northern part of the P-W terrane. The



EXPLANATION

Sedimentary rocks of Paleozoic age

MIDDLE PROTEROZOIC

Yar — Anorogenic igneous rocks (1,470–1,500 Ma)

EARLY PROTEROZOIC

Xq — Quartzite—In part older than unit Xr

Xr — Rhyolite (≈1,760 Ma)

Xga — Granitoid rocks (≈1,760 Ma)

Xgr — Alkali-feldspar granite (≈1,835 Ma)

Xgt — Granite to granodiorite (≈1,836 ± 15 Ma)

Xgg — Tonalite-granodiorite and gneissic granodiorite (1,855–1,870 Ma)

Xmv — Volcanic and sparse sedimentary rocks of oceanic arc affinity—Greenschist and lower amphibolite metamorphic grade (1,840–1,889 Ma)

EARLY PROTEROZOIC AND ARCHEAN

XWr — Volcanic and sedimentary rocks (≈1,860 Ma) overlying Archean (≈2,800 Ma) basement; intruded by tonalite-granite (1,840–1,892 Ma)

ARCHEAN

Wgn — Gneiss, migmatite, and amphibolite (≈2,800 Ma)

Contact

High-angle fault—Dashed where inferred

High-angle normal fault—Bar and ball on downthrown side

Thrust fault—Dashed where inferred; sawteeth on upper plate

(C) Locality referred to in text

Figure 26. Generalized geologic map of Wisconsin magmatic terranes, Penokean orogen. Area is south of that of figure 22. Unit Xr (≈1,760 Ma) overlaps volcanic-plutonic sequences. D, Dunbar; H, Holbrook; GB, Green Bay; M, Marshfield; P, Pembine; R, Rhinelander; W, Wausau; WR, Wisconsin Rapids. Modified from Sims and others (1989) and Sims (1992).

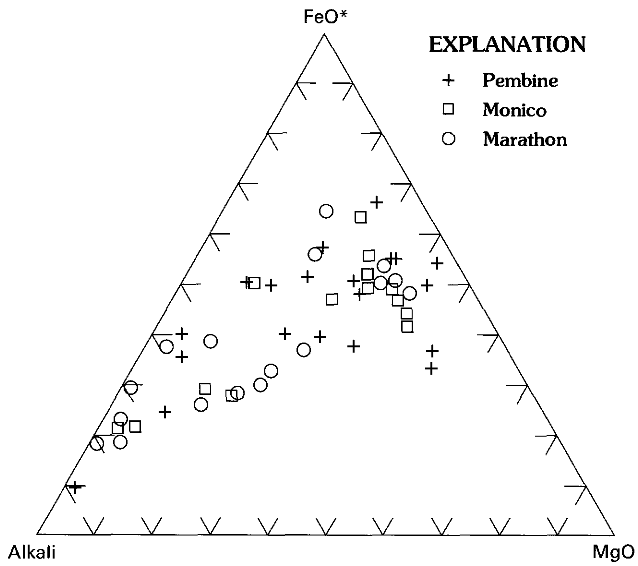


Figure 27. AFM diagram for volcanic rocks from Pembine, Monico, and Marathon County areas, Pembine-Wausau terrane, northern Wisconsin. Pembine is in northeastern part, Monico in central part, and Marathon County in southern part of the terrane. Modified from Sims and others, 1989.

McGrath–Little Falls terrane in east-central Minnesota (fig. 24; unit XAi, pl. 1) possibly is partially equivalent to the Wisconsin magmatic terranes.

PEMBINE-WAUSAU TERRANE

The P-W terrane is a 150-km-wide, apparently wedge shaped domain of volcanic and plutonic rocks (fig. 26) that is made up of two volcanic successions, an older one deposited in the interval 1,889–1,860 Ma and a younger, more restricted but poorly delineated one deposited between 1,845 and 1,835 Ma. The younger succession is inferred to unconformably overlie the older one in Marathon County and vicinity (LaBerge and Myers, 1984) in the southern part of the P-W terrane (vicinity of Wausau (W), fig. 26). The volcanic rocks are intruded by coeval syntectonic plutons (1,870–1,840 Ma) and by younger post-tectonic intrusions, 1,835 Ma and ≈1,760 Ma.

VOLCANIC ROCKS: 1,889–1,860 MA SUCCESSION

The older volcanic succession in the P-W terrane consists mainly of metamorphosed basalt, andesite, dacite, and rhyolite flows and pyroclastic rocks. In northeastern Wisconsin (Sims, 1990b; Sims and others, 1992), two depositional cycles are recognized. The older rocks are tholeiitic basalt and basaltic andesite, formally named the Quinnesec Formation, which are a minimum of 1,889 Ma, and probably older. These rocks show limited FeO enrichment (fig. 27) and low TiO_2, and they are relatively depleted in high-field-strength elements. Chondrite-normalized rare-earth element (REE) patterns show moderate to extreme light rare-earth element depletion (Sims and others, 1989, fig. 3*A*). The tholeiitic basalts are similar compositionally to some recent back-arc basin basalts and island-arc tholeiites. The strongly LREE-depleted basalts are particularly similar to some ophiolitic basalts, including those from the Oman ophiolite (Pearce and others, 1981) and the Troodos ophiolite (Kay and Senechal, 1976), both of which are now interpreted as having evolved in island arcs (Moores and others, 1984; Schmincke and others, 1983; Pearce and others, 1981). The basalts in northeastern Wisconsin are intruded by sheeted dikes, serpentinites, and plagiorhyolite (Schulz, 1987), suggesting that these rocks are a dismembered ophiolite.

The presumably younger calc-alkaline volcanic rocks, as represented in northeastern Wisconsin (Sims, Schulz, and others, 1993), range in composition from andesite to rhyolite; they have major- and trace-element compositions similar to average K_2O calc-alkalic andesitic suites. They have moderately enriched heavy rare-earth element patterns and increasingly negative Eu anomalies with increasing SiO_2 content (Sims and others, 1989, fig. 3*B*).

Volcanic rocks in the central part of the P-W terrane (locality D, fig. 26) constitute a bimodal suite; pillowed flows of high-aluminum basalt and low-SiO_2 andesite compositions are intercalated with dacite to rhyolite tuffs and porphyries. Both the mafic and felsic volcanic rocks show enriched LREE with flat HREE patterns and relative depletion in tantalum (Sims and others, 1989, fig. 4). These rocks contain at least three major copper-zinc massive sulfide deposits (Sims, 1987 and references therein), and are compositionally similar to the bimodal calc-alkalic volcanic rocks that host the Kuroko massive sulfide deposits of Japan (Dudas and others, 1983).

VOLCANIC ROCKS: 1,845–1,835 MA SUCCESSION

A younger succession of calc-alkalic andesite, dacite, and rhyolite that is distinguished from older tholeiitic basalt in the Marathon County area (Sims, 1990a) primarily by having a different structural style and lower metamorphic grade (LaBerge and Myers, 1984) is poorly dated and somewhat enigmatic. The volcanic rocks have more enriched LREE and large-ion lithophile element abundances than the calc-alkalic rocks in northeastern Wisconsin, which suggests a geographical compositional zonation within the P-W terrane. The geographical zonation, which is common for subduction-related magmatic sequences (Saunders and others, 1980), could be caused by more crustal involvement than to the north or by differences in the distance from the subduction zone (Niagara fault zone). Existing Sm-Nd data (Barovich and others, 1989), however, do not suggest involvement

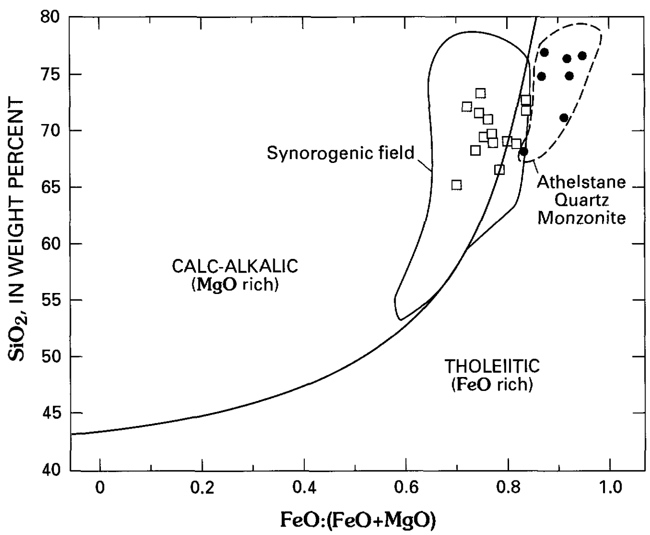


Figure 28. SiO_2 (weight percent) versus FeO:(FeO+MgO) diagram for the syntectonic granodiorite suite (open square) and the post-tectonic Athelstane Quartz Monzonite (solid dot), Pembine-Wausau terrane. Synorogenic field of Anderson (1983) shown for comparison.

of Archean basement rocks in this part of the P-W terrane. Further studies of the zoning are needed.

GRANITOID ROCKS

Granitoid rocks in the P-W terrane are mainly granodiorite but range from tonalite and gabbro to alkali granite. They can be grouped into two broad classes: (1) a syntectonic granodiorite suite and (2) a post-tectonic suite consisting of the ≈1,835 Ma Athelstane Quartz Monzonite of the Athelstane batholith, small bodies of an 1,835 Ma red alkali-feldspar granite, and small bodies of a ≈1,760 Ma granodiorite-granite group (Sims, Schulz, and others, 1993).

SYNTECTONIC GRANODIORITE SUITE

The syntectonic granodiorite suite is spatially associated with the major (1,860-1,889 Ma) sequence of volcanic rocks (most of unit Xmv, fig. 26) in the P-W terrane. The rocks of the syntectonic granodiorite suite are foliated and are mainly included in unit Xgg in figure 26; the suite is calcic to calc-alkalic and has intermediate SiO_2 concentrations, intermediate to high Al_2O_3 (12.4–16.4 percent), high Sr concentrations, and low Rb, Nb, and Y concentrations. These rocks are compositionally similar to the well-studied Newingham Tonalite in the Dunbar, Wis., area (Sims and others, 1992), which is a volcanic-arc granite about 1,860 Ma. On a SiO_2 versus FeO:(FeO+MgO) diagram (fig. 28), rocks of the syntectonic granodiorite suite plot mainly in the calc-alkalic (MgO-rich) field, and lie within the synorogenic field of Anderson (1983) for Penokean granitoid rocks in Wisconsin. The synorogenic field is virtually distinct from the field for the Athelstane Quartz Monzonite and the 1,835 Ma alkali-feldspar granite, indicating distinctly different magmas for the syntectonic and post-tectonic rocks. Tonalite and granodiorite exposed in central Wisconsin (unit Xgt, fig. 26) are included in the granodiorite suite, but they are distinctly younger than most of the granitoids exposed farther north.

Except for granodiorite-tonalite bodies in central Wisconsin (unit Xgt, fig. 26), the syntectonic granodiorite suite is interpreted as products of southward subduction of oceanic crust prior to collision of the north-verging P-W terrane with the continental margin to the north (Sims, Schulz, and others, 1993), presumably as a partial melt of basalt. The Dunbar Gneiss, in northeastern Wisconsin, and presumably other gneissic granitoids adjacent to the Niagara fault zone, however, are more evolved than the Newingham Tonalite and other volcanic-arc granitoids (Sims and others, 1992). These rocks are characterized by relatively high concentrations of K, Rb, Ba, Th, Nb, Ta, and LREE, and are interpreted as syncollisional granites. The younger granodiorite-tonalite bodies in central Wisconsin presumably formed during north-dipping subduction of oceanic crust that terminated with collision of the Marshfield (minicontinent) terrane with the P-W terrane at about 1,840 Ma (Sims, Schulz, and others, 1993). These rocks are somewhat more evolved than the Newingham Tonalite and other volcanic-arc granodioritic rocks farther to the north.

POST-TECTONIC SUITE

The post-tectonic suite comprises those plutonic rocks emplaced subsequent to ≈1,850 Ma, and consists of the ≈1,835 Ma Athelstane Quartz Monzonite, 1,835 Ma alkali-feldspar granite, and a ≈1,760 Ma granodiorite-granite group (Sims, Schulz, and others, 1993).

ATHELSTANE QUARTZ MONZONITE

The Athelstane Quartz Monzonite constitutes most of the 900 km^2 Athelstane batholith in northeastern Wisconsin (unit Xgt, fig. 26; Sims, 1992). It is a massive coarse-grained granite to granodiorite except along the north margin of the batholith, where it has a steep stretching lineation resulting from uplift of the batholith subsequent to crystallization of the magma. One sample has a U-Pb zircon upper intercept age of 1,836±15 Ma (Sims, 1990b). Granite near Cherokee (unit Xgt, fig. 26), southwest of Wausau, is considered correlative with the Athelstane Quartz Monzonite. The Athelstane and granite near Cherokee are mildly peraluminous, and have high SiO_2 (68–77 percent), intermediate Al_2O_3 (12–15 percent), and Na:K $\leq$1. These rocks are more evolved chemically than the syntectonic granodiorite suite; they are strongly enriched in iron, and plot in the anorogenic (FeO-rich) field in the FeO:(FeO+MgO) diagram (fig. 28).

1,835 MA ALKALI-FELDSPAR GRANITE

The 1,835 Ma alkali-feldspar granite (unit Xgr, fig. 26) crops out in the southern part of the P-W terrane and truncates the Eau Pleine shear zone (fig. 26). These rocks form small, separate plutons. The rocks lack a penetrative foliation but contain abundant brittle fractures and show extensive retrogressive metamorphism. The granite is peraluminous, alkali-calcic, and Fe rich, and has very low Sr and high U, high FeO:(FeO+MgO) (O >0.80), Na_2O:K_2O (<1.0), Ba:Sr (>3.0), and K:Rb (>300), and large negative Eu anomalies (Sims and others, 1989, fig. 10). On a SiO_2 versus FeO:(FeO+MgO) diagram (fig. 28), these rocks plot with the Athelstane Quartz Monzonite in the anorogenic field.

1,760 MA GRANODIORITE-GRANITE GROUP

The 1,760 Ma granodiorite-granite group crops out as scattered small plutons (unit Xga, fig. 26) across northern Wisconsin. The rocks lack a penetrative foliation but are cut by brittle-ductile shears. Zircon ages average ≈1,760 Ma but range from 1,739±8 Ma to 1,773±8 Ma (Sims and others, 1989). The rocks are coeval with the anorogenic rhyolite-granite terrane in south-central Wisconsin (Smith, 1983), but differ from these rocks in being more variable in composition and in having somewhat lower FeO:(FeO+MgO) and higher Sr.

MARSHFIELD TERRANE

The Marshfield terrane, south of the Eau Pleine shear zone, differs from the P-W terrane mainly in having an extensive Archean basement to the volcanic rocks (fig. 26; Sims, 1990a). Also, a widespread superposed steep foliation and lineation and associated mylonite, developed some time between 1,860 and 1,835 Ma, has largely obliterated original structures in both the Archean and Early Proterozoic rocks. Because of this structural overprint and the generally sparse exposures, knowledge of the terrane is meager.

VOLCANIC ROCKS

The Marshfield terrane consists of apparent erosional remnants of an approximately 1,860 Ma mafic to felsic volcanic succession overlying at least a partial Archean basement (Sims, 1990c). The major exposures comprise the Milladore Volcanic Complex (locality A, fig. 26), an interlayered sequence of metamorphosed felsic to mafic volcanic rocks, dacite porphyry, and sedimentary rocks (impure quartzite, ferruginous chert, conglomerate, and carbonaceous argillite). These rocks are intruded by small bodies of metagabbro and metadiorite and by a large (35 km^2) pluton of foliated tonalite. A smaller area of volcanic and sedimentary rocks is exposed in the Eau Claire River valley (locality B, fig. 26). The rocks in both areas are mainly metamorphosed to upper greenschist (garnet) grade.

GRANITOID ROCKS

Granitoid rocks of Early Proterozoic age in the Marshfield terrane are mainly tonalite-granodiorite and alkali-feldspar granite, but range from gabbro-diorite to alkali granite (Sims, Schulz, and others, 1993). The rocks can be grouped on the basis of composition, structure, and age into two classes: (1) a syntectonic tonalite-granodiorite suite and (2) 1,835 Ma post-tectonic alkali-feldspar granite.

SYNTECTONIC TONALITE-GRANODIORITE SUITE

The tonalite-granodiorite suite crops out as small bodies that intrude Archean gneiss in the area east of Wisconsin Rapids (fig. 26) and as a modest-sized pluton that cuts ≈1,860 Ma metavolcanic rocks of the Milladore Volcanic Complex (locality A, fig. 26). Anderson (1972) described the field relations and petrology of the bodies that cut Archean gneisses. Rocks of the tonalite-granodiorite suite are presumably coeval with the ≈1,860 Ma Milladore Volcanic Complex, yet one sample of foliated diorite has a U-Pb zircon upper intercept age of 1,892±9 Ma (Sims and others, 1989; Sims, Schulz, and others, 1993), suggesting that the volcanic rocks are in part at least older than ≈1,860 Ma.

Granodiorite to tonalite rocks in the area range from hornblende-biotite diorite to biotite trondhjemite and are calc-alkalic to alkali-calcic (Sims, Schulz, and others, 1993). The rocks are mostly metaluminous but include peraluminous varieties for silica values greater than 65 weight percent. Anderson and Cullers (1987) showed that these plutons are analogs of Holocene orogenic andesite-dacite suites; a small amount of Archean material possibly was incorporated during tonalite emplacement.

A small body of gneissic granite near Neillsville, along the Black River (locality C, fig. 26), has a comparably old age (1,871±5 Ma) to the tonalite-granodiorite suite but is more evolved chemically (Sims, Schulz, and others, 1993).

1,835 MA ALKALI-FELDSPAR GRANITE

The post-tectonic red alkali-feldspar granite is similar compositionally to the alkali-feldspar granite in the P-W terrane, and the rocks in both terranes are presumed to be cogenetic (Sims, Schulz, and others, 1993). The REE patterns are distinctive, exhibiting large negative Eu anomalies and a relatively flat HREE slope (Sims, Schulz, and others, 1993, fig. 14). The granite mainly plots in the within-plate field on a Rb versus Y+Nb diagram.

MAGMATIC TERRANE IN EAST-CENTRAL MINNESOTA

The magmatic terrane in eastern Minnesota (unit XAi, pl. 1) is a rather poorly exposed complex of granitoid rocks intruded into metamorphic rocks of probably both Archean and Proterozoic ages (fig. 24; Southwick and others, 1988). The relationship of this terrane to the terranes east of the Midcontinent rift system is uncertain, but Southwick and Morey (1991) have tentatively correlated it with the Wisconsin magmatic terranes because it contains abundant calc-alkalic syntectonic to post-tectonic granitoid rocks, as do the Wisconsin magmatic terranes (fig. 11). Following the earlier structural interpretation of Southwick and others (1988), Southwick and Morey (1991) included the McGrath gneiss dome (age ≈2,700 Ma; Stuckless and Goldich, 1972) and its mantling sedimentary rocks of Early Proterozoic age (Denham Formation) in the magmatic terrane (also see Chandler and Southwick, 1990). They extended the magmatic terrane (McGrath–Little Falls terrane, fig. 24) northward to the Malmo discontinuity, which they tentatively correlated with the Niagara (suture) fault zone east of the Midcontinent rift system.

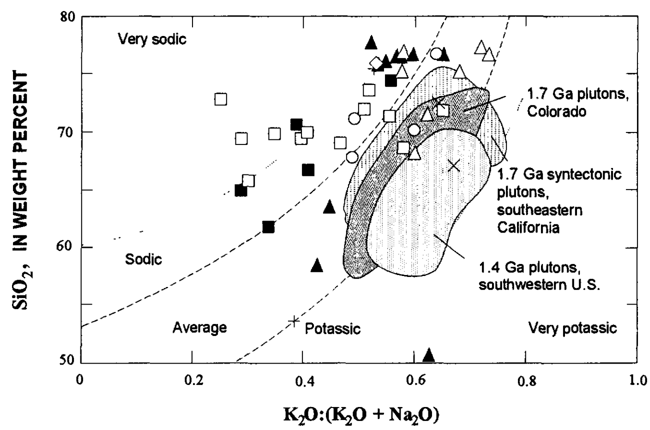
The syntectonic granitoid rocks in the magmatic terrane are foliated granodiorite to granite bodies that are interpreted to occupy cores of domal structures in migmatite (Hillman Migmatite of Morey, 1978). Southwick and others (1988) concluded that these plutons are syntectonic to late-tectonic diapirs. One intrusive body (Freedhem Granodiorite) has a U-Pb zircon age of 1,868±3 Ma (Goldich and Fischer, 1986), which is coeval with the syntectonic granitoid suite in the Pembine-Wausau arc terrane in Wisconsin (Sims and others, 1989).

Intrusive rocks in the northeasternmost part of unit XAi (pl. 1), the Isle and Warman Granites of Morey (1978), are post-tectonic granite and granodiorite of ≈1,770 Ma (Spencer and Hanson, 1984). These rocks are interpreted (Southwick and others, 1988) to intrude the Denham Formation on the south side of the McGrath gneiss dome. The post-tectonic granitoid rocks are similar in age and composition to the post-tectonic intrusions of ≈1,760 Ma in the P-W (Pembine-Wausau) terrane in Wisconsin (Sims and others, 1989).

A migmatite of uncertain age (Hillman Migmatite of Morey, 1978) occupies several tens of square kilometers in the unit XAi, and is the host for the syntectonic granitoid bodies (Southwick and others, 1988; Tasker, 1983). The rock is a gray foliated biotite-garnet-cordierite schist, hornblende schist, and biotite-feldspar-quartz granofels migmatized by tonalitic neosome. The tonalitic neosome is intruded by massive tonalite that has a U-Pb zircon age of 1,869±5 Ma (Goldich and Fischer, 1986). The age of the paleosome has not been determined.

EVOLUTION

The Wisconsin magmatic terranes evolved through a complex series of tectonic events following extension and breakup of the Early Proterozoic continental margin. Southward subduction led to the development of a major volcanic arc represented by the 1,889–1,860 Ma rocks in the P-W terrane. The arc rocks apparently accumulated on oceanic crust, as suggested by a probable dismembered ophiolite in the vicinity of the Niagara fault zone. The basaltic rocks could have formed in a closed back-arc basin, such as the Holocene Lau basin in the Pacific Ocean, or as island arcs. The bimodal volcanic rocks in the central part of the P-W terrane could have formed in a back-arc basin. Arc volcanism was accompanied by intrusive bodies of tonalite and granodiorite of volcanic arc genesis. Continuing subduction led to



EXPLANATION

Marshfield terrane

▲ 1,835 Ma alkali-feldspar granite
■ Syntectonic granodiorite
+ Uncertain affinity

Pembine-Wausau terrane

△ Athelstane Quartz Monzonite
□ Syntectonic granodiorite
× Uncertain affinity
○ 1,760 Ma granodiorite-granite
◇ 1,835 Ma alkali-feldspar granite

Figure 29. SiO_2 (weight percent) versus $K_2O:(K_2O+Na_2O)$ diagram showing comparison of compositions of plutonic rocks in Pembine-Wausau and Marshfield terranes with those in Proterozoic terranes in Colorado, southeast California, and southwest United States. From Sims, Schulz, and others, 1993, fig. 18.

collision of the arc complex with the continent at about 1,850 Ma and the development of collision-zone intrusive bodies, such as the Dunbar Gneiss in northeastern Wisconsin. The collision-zone intrusions are relatively enriched in high-field-strength element abundances, particularly tantalum and niobium. Such a composition is compatible with these granitoid rocks having been derived by partially melting mantle-derived basalts that have within-plate compositions and that are older than the collision-zone intrusions (for example, Early Proterozoic basalts of the Marquette Range Supergroup). Partial melting would have occurred during collision and overthrusting of the P-W terrane onto the continental margin.

The 1,845–1,835 Ma felsic volcanic rocks erupted in the southern part of the P-W terrane possibly were generated above a north-dipping subduction zone during the closure of an ocean basin between the accreted P-W terrane and a minicontinent (Marshfield terrane) to the south. Subsequent collision of the two terranes along the Eau Pleine shear zone possibly caused the widespread ductile deformation in rocks of the Marshfield terrane, which preceded emplacement of the post-tectonic (≈1,835 Ma) alkali granite.

The emplacement of the scattered 1,760 Ma granitoid bodies in the northern part of the P-W terrane probably resulted from continent-continent or continent-arc collision to the south of the Lake Superior region that was contemporaneous with the 1,760 Ma anorogenic magmatism in southern Wisconsin. Accretion of the Central Plains orogen (Sims and Peterman, 1986) to the North American craton could have triggered this magmatism.

The plutonic rocks of the P-W and Marshfield terranes are similar in composition, and they record changing types of magmatism over time. Prior to ≈1,850 Ma, intrusive rocks are calcic to calc-alkalic, magnesium rich, sodic to very sodic, high-strontium granodiorite to trondhjemite typical of immature island or continental arcs (Sims, Schulz, and others, 1993). Gneissic granitoids emplaced in the proximity of the Niagara fault zone (fig. 26) are more evolved chemically than the volcanic-arc granitoids, and are interpreted as collisional-zone intrusive bodies (Sims and others, 1992). Rocks younger than ≈1,850 Ma are characteristically alkali-calcic, iron rich to very iron rich, sodic to average granodiorite to granite typical of more mature island arcs and somewhat evolved continental arcs. The post-tectonic 1,760 Ma plutons are the most chemically evolved; they have the highest rubidium and thorium concentrations, but they do not differ much from the 1,835 Ma plutons, as they are alkali-calcic, iron rich to average, sodic to average granodiorite to granite (Sims, Schulz, and others, 1993).

A distinctive and as yet unexplained feature of the igneous rocks of the P-W and Marshfield terranes is the overall sodic nature of the magmatism, which continued from ≈1,890 to ≈1,500 Ma in northern Wisconsin (fig. 29). Whereas other orogenic terranes in the Western United States typically evolved to potassic or even highly potassic magmatism over time, the Wisconsin terranes remained sodic to average. Perhaps further studies will lead to an explanation of this anomaly.

EARLY AND MIDDLE PROTEROZOIC INTRACRATONIC ROCKS

By P.K. Sims

A rhyolite-granite (≈1,760 Ma) terrane in southern Wisconsin, quartzite, and epizonal granite bodies, including the Wolf River batholith (≈1,470 Ma), comprise major intracratonic rock units in the Lake Superior region (pl. 1, Yw, Ygr, Xrg, Xq). They postdate the Penokean orogeny, and overlie and (or) intrude Archean and Early Proterozoic rocks in the orogen.

1,760 MA RHYOLITE-GRANITE TERRANE

Rhyolite and granite ranging in age from 1,773±8 Ma to 1,739±8 Ma (Sims and others, 1989, table 1) crop out sparsely in the Fox River valley and tributaries, in southern Wisconsin (Sims, 1992). The rocks are interpreted from magnetic anomaly maps to occupy an area of about 10,000 km^2 (pl. 1). The rocks compose a northeast-trending basin or basins of probable extensional origin, the northwest margin of which is rather well delineated by faults disclosed by aeromagnetic data. The southern extent of these rocks is uncertain. The rhyolite comprises interbedded ash-flow tuff and volcaniclastic sedimentary rocks of both peraluminous and metaluminous suites (Anderson and others, 1980; Smith, 1983). Associated cogenetic granites are granophyric and probably intruded their own volcanic cover. An undated diorite body and associated basaltic dikes locally intrude the rhyolite. The rhyolitic and associated rocks are weakly altered to retrogressive assemblages, particularly hematite and chlorite, and are mildly deformed on northeast-trending fold axes. The peraluminous and metaluminous suites are possibly comagmatic, and have been interpreted as having been derived from partial melting of separate sources of intermediate composition (Smith, 1983). The diorite and associated basalt probably were derived by partial melting of a mantle source and should have provided the heat source for partial melting of the crust. No known mineral deposits are associated with this anorogenic suite, but the rocks are believed to have some potential for Olympic Dam-type deposits (Sims, 1990f).

QUARTZITE

Quartz-arenitic red-bed sequences (pl. 1, Xq) crop out sporadically in the southern part of the Lake Superior region as isolated bodies that include the Sioux, Barron, Flambeau, McCaslin, Baraboo, and Waterloo Quartzites, as well as several other smaller, unnamed sequences. Because of their similarity in appearance, the red-bed sequences have been presumed to be broadly correlative (Dott and Dalziel, 1972). The prevailing model assumed that all the red-bed sequences were once part of a continuous blanket of sand deposited in a shallow-water nearshore environment on an east-trending southward-sloping continental shelf (Dott and Dalziel, 1972). Subsequently, Dott (1983) modified the sedimentological model, suggesting that the red-bed sequences could have been deposited as braided fluvial deposits formed on a sandy coastal plain several kilometers wide that subsequently was submerged during a shallow marine transgression. Although realizing that individual sequences are not necessarily correlative, Dott (1983) proposed that the red beds formed during the time interval 1,750–1,450 Ma, which he termed the "Baraboo interval." Later, Greenberg and Brown (1984) redefined the Baraboo interval as that period of time characterized by anorogenic igneous and sedimentary activity between about 1,760 Ma and 1,500 Ma.

Subsequent studies have shown that indeed all the red-bed sequences are not strictly correlative. In a broad paleomagnetic study, Chandler and Morey (1992) showed that the Sioux Quartzite in southwestern Minnesota formed in the approximate interval 1,700–1,650 Ma; the Sioux paleopole cannot be distinguished at the 95 percent confidence level from that of the Barron Quartzite in northwestern Wisconsin. On the basis of existing apparent polar wander data, Chandler and Morey (1992) proposed that the Baraboo Quartzite could be as much as 100 m.y. older than the Sioux and Barron. An older age for the Baraboo Quartzite is supported by structural studies of LaBerge and others (1991). These writers proposed on indirect evidence that the Baraboo and other smaller quartzite bodies in central and southern Wisconsin, south of the Eau Pleine shear zone, are 1,860 Ma or older. They proposed that these quartzite bodies are allochthonous, having been transported southward by a major south-vergent fold-thrust system related to the Eau Pleine shear zone. Earlier, Sims and others (1988) showed on indirect evidence that the McCaslin Quartzite in northeastern Wisconsin is older than 1,812 Ma. A deformed conglomerate (Baldwin Conglomerate) within the Mountain shear zone (Sims, Klasner, and Peterman, 1991) contains quartzite clasts that almost certainly were derived by erosion from the nearby McCaslin Quartzite. The conglomerate is older than an undeformed intrusion, the Hines Quartz Diorite (1,812 Ma), which was emplaced into the Mountain shear zone at or near the termination of shearing.

The cumulative evidence that quartzite bodies in the Lake Superior region differ in age, and perhaps comprise two temporally distinct red-bed sequences—one late Penokean in age and the other post-1,760 Ma—indicates that the rocks of "Baraboo interval" have no paleogeographic continuity, as stated previously by Chandler and Morey (1992).

WOLF RIVER BATHOLITH

The Wolf River batholith of central Wisconsin is one of the older intrusions of the 1.4–1.5 Ga transcontinental anorogenic province of North America (Anderson, 1983). The batholith has an age of ≈1.47 Ga (recalculated by Z.E. Peterman from Van Schmus and others, 1975). Associated anorogenic intrusions of approximately the same age include the Wausau and Stettin plutons in the Wausau area of central Wisconsin (Sims, 1990a). The petrology and geochemistry of the Wolf River batholith have been studied in some detail (Anderson and Cullers, 1978; Anderson, 1980, 1983, 1993). This description is mainly taken from these reports.

The Wolf River batholith has many features in common with other Middle Proterozoic anorogenic granites in the transcontinental anorogenic province, but also has some unique features. Similarities include (1) inclusions of anorthosite and charnockitic rocks, (2) high alkali-feldspar content, (3) several textural varieties of rapakivi granite, (4) accessory allanite, apatite, zircon, and fluorite±sphene, and (5) enrichment in K, Rb, Ba, REE, and U (Anderson, 1993). It is unique in (1) comprising nine petrographically distinct granitic plutons, (2) being ilmenite bearing, (3) having crystallized at low fO_2 (approximately QFM), and (4) that the younger plutons are increasingly less felsic (> Ca, Mg, Al, Sr, and < REE).

The Wolf River batholith has an exposed area of about 10,000 km^2 and comprises five major granite plutons in addition to a central core of anorthosite and two peripheral

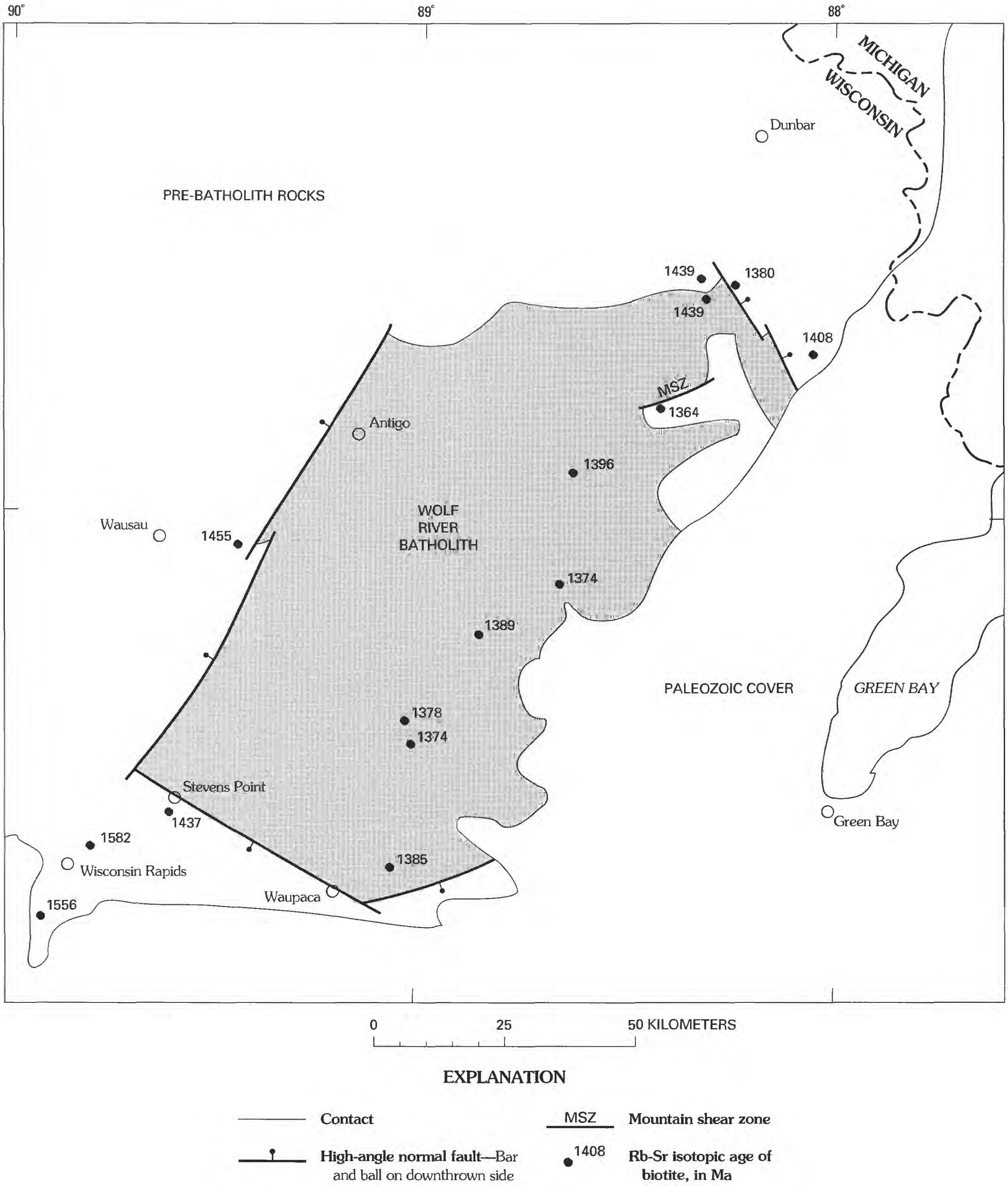


Figure 30. Rb-Sr biotite ages in Wolf River batholith and surrounding country rocks, Wisconsin. Ages indicate that batholith was differentially uplifted relative to country rock about 100 m.y. after emplacement. Eastern part of batholith is covered by sedimentary rocks of Cambrian age.

bodies of mangerite. The eastern part is covered by Paleozoic sedimentary rocks (Sims, 1992). The oldest and largest intrusion is the Wolf River Granite, a coarse-grained rapakivi granite. Intrusive into the Wolf River Granite is the second largest pluton, the Red River Granite, as well as monzogranite dikes termed Wiborgite porphyry. The southeast margin of the batholith is the Waupaca Granite (Sims, 1990a), a coarse-grained Wiborgite-type granite. The northeastern part of the batholith includes the Belongia Granite, which is a red to pink granite having both a coarse and a fine facies (Sims, 1990b) and is probably the youngest intrusion in the batholith. Another late intrusion is the High Falls Granite, which forms the northeast margin of the batholith (Sims, 1990b). The granite plutons in the northeastern part of the batholith intrude the Hager Formation, which consists of a substantial body of rhyolite and a porphyritic subvolcanic member (Sims, 1990b). Accordingly, the northeastern part of the batholith is considered the highest part of the batholith, which intruded its volcanic cover.

Exposed margins of the batholith are bounded by high-angle faults. The faults and adjacent granitoid rocks are marked by mylonite and (or) cataclastic rocks, indicating movement on the structures after emplacement of the intrusive rocks. That uplift of the batholith has indeed taken place subsequent to crystallization of the igneous rocks is indicated by Rb-Sr biotite ages on samples of rocks within the Wolf River batholith (fig. 30). The mean biotite age of six samples of igneous rocks is 1,383±7 Ma, which is nearly 100 m.y. younger than the crystallization age of the batholith. These low biotite ages are interpreted as registering closure due to cooling below the 300 °C isotherm as a consequence of buoyant uplift and rapid erosion of rocks within the batholith.

The existence of remnants of volcanic rocks in the northeastern part of the Wolf River batholith indicates that the batholith intruded very high in the crust. The differentiated plutons (Belongia Granite) have compositions near the granite minima corresponding to a pressure of 0.5 to 1.0 kb or a depth of 1.9 to 3.8 km (Anderson and Cullers, 1978). Epizonal megascopic features, high crystallization temperatures (650–840 °C from feldspar thermometry), hornblende with low total alumina (as low as 7.41 wt. percent Al_2O_3 toward north margin), and low water fugacity are also suggestive of high-level emplacement (Anderson, 1980).

Anderson and Cullers (1978) demonstrated that the granitic magmas probably formed from partial melting of a crustal igneous source of tonalitic to granodioritic composition at a crustal depth in the range 27–36 km. Mangerite and (or) anorthosite may have provided at least part of the heat source necessary to generate a granitic melt. A low $^{87}Sr/^{86}Sr$ ratio of 0.7048±0.0017 (Van Schmus and others, 1975) indicates that the source is nonradiogenic and possibly Early Proterozoic (Penokean) in age; Sm/Nd isotopic data (Nelson and De Paolo, 1985) confirmed this model. The batholith appears to have been emplaced at the intersection of at least three sets of deep-seated crustal faults (pl. 1).

The Wolf River batholith is overlain by the Wisconsin gravity minimum, a 50,000 km^2 anomaly that is one of the most prominent isostatic residual anomalies in the continental United States (Allen and Hinze, 1992). Modeling by Allen and Hinze indicated that the source of the gravity low is the Wolf River batholith, and that the buried extent of the batholith is far greater than suggested by the exposed area. The model suggests that the bottom of the batholith lies at a depth of about 10 km and that its top is overlain by, at most, 4 km of older Precambrian rocks.

MIDDLE PROTEROZOIC MIDCONTINENT RIFT SYSTEM

By William F. Cannon *and* Suzanne W. Nicholson

The Midcontinent rift system (MRS) formed during a brief (15 m.y.) period of extension in which the Archean and Early Proterozoic crust was nearly, or perhaps totally, separated concomitantly with the formation of a great thickness of basaltic volcanic rocks. An ensuing, somewhat longer period of thermal subsidence resulted in deposition of a sequence of mostly fluvial clastic rocks in a flexural basin more or less centered on the axis of earlier rifting. Although these sedimentary rocks were not deposited in an extensional regime, together with the volcanic rocks they are generally considered to be products of the Midcontinent rift. The volcanic and sedimentary rocks together comprise the Keweenawan Supergroup.

Rocks of the MRS are exposed discontinuously around Lake Superior (fig. 31), generally in thick monoclinal successions that dip toward the rift axis beneath the lake. Geologic structures, mostly reverse faults, repeat the section in some areas and slightly complicate the otherwise simple

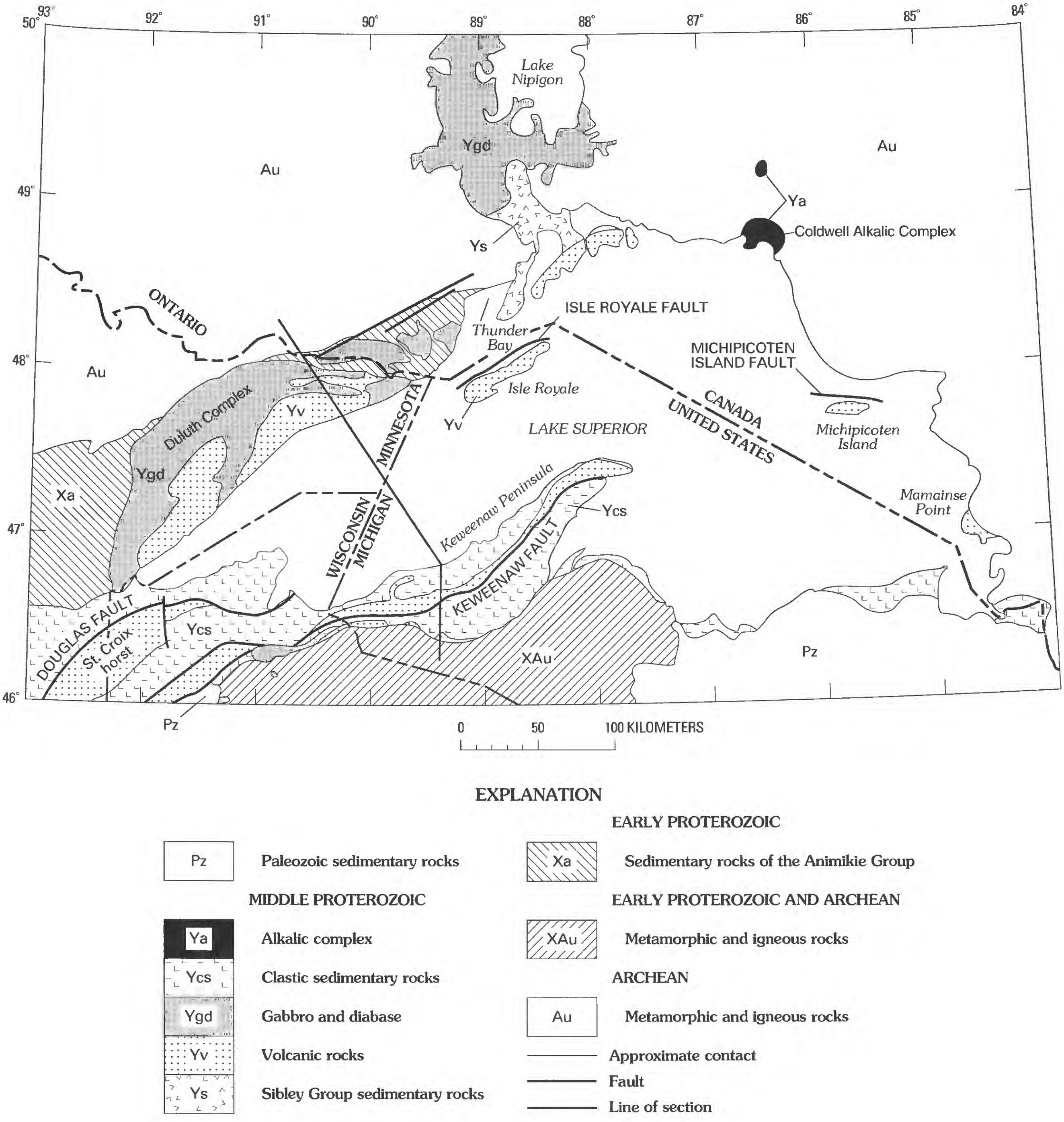


Figure 31. Generalized geology of the Lake Superior region showing major volcanic and sedimentary units of the Keweenawan Supergroup and rift-related intrusive rocks. Heavy line shows approximate location of cross section in figure 33.

structure. Details of the geology of the exposed rocks are presented in a series of summary papers in Wold and Hinze (1982), which are in part updated by more recent summaries by Green (1983, 1989), Massey (1983), Brannon (1984), Paces (1988), Klewin and Berg (1990, 1991), Van Schmus and Hinze (1985), and Weiblen (1993). The reader is referred to these papers to supplement the brief geologic description given here.

Keweenawan rocks of the MRS are also known from geophysical data and drilling to occur in the subsurface, beneath Paleozoic rocks, southwestward as far as Kansas, and southeastward to southern Michigan.

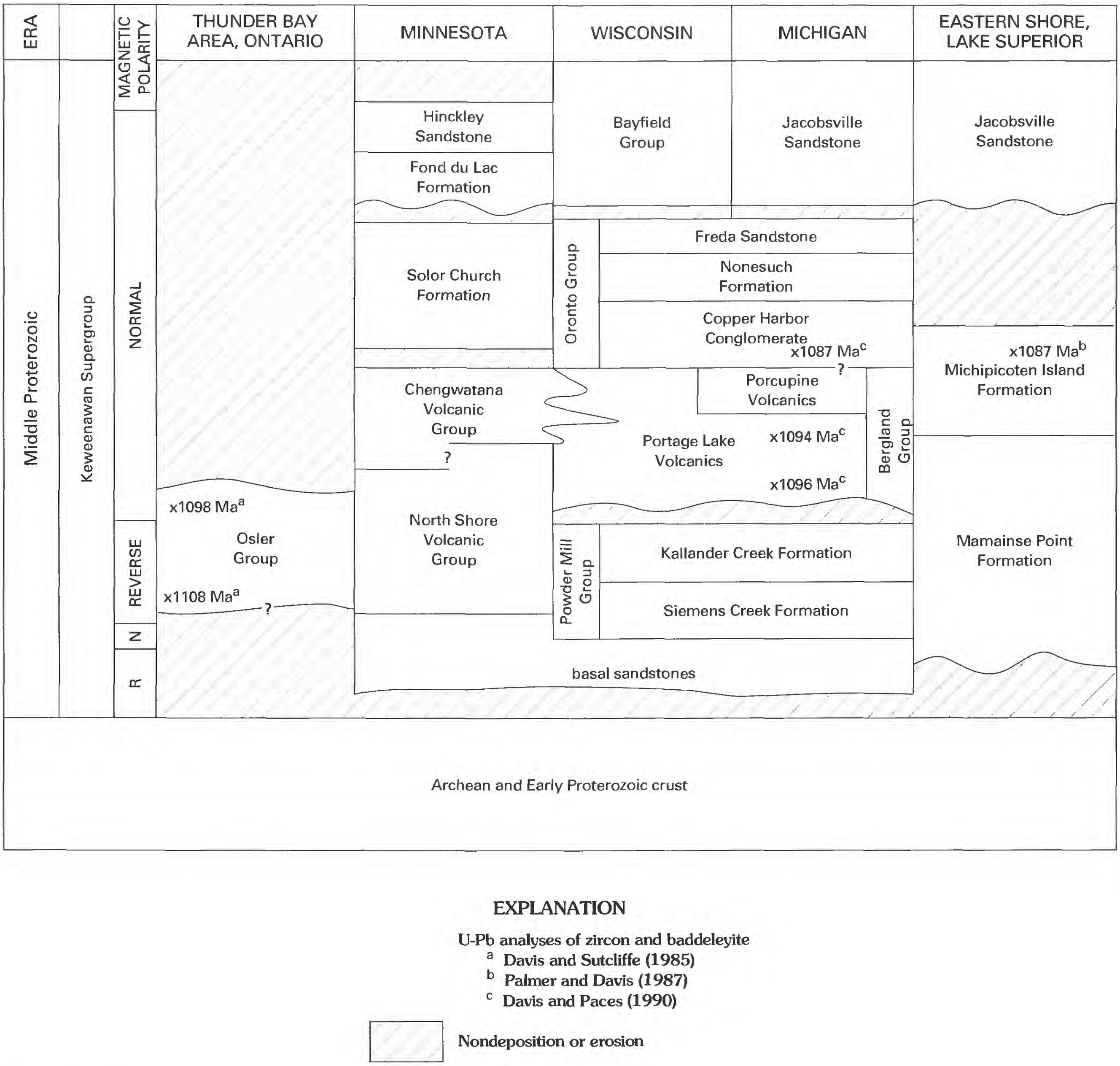


Figure 32. Stratigraphic correlation diagram for Middle Proterozoic rocks in the Lake Superior region, from Cannon and Nicholson (1992). Wavy line, unconformity; queried where uncertain. Contact queried where uncertain.

PREVOLCANIC SEDIMENTARY ROCKS

Relatively thin (mostly less than 100 m) units of mature fluvial quartz sandstones are common in the western part of the Lake Superior region (Ojakangas and Morey, 1982), and were the first rocks deposited in the rift basin (fig. 32). They lie unconformably on Early Proterozoic and Archean rocks. Paleocurrent directions are mostly toward or parallel to the axis of the rift basin, indicating that subsidence at this time was more or less centered on the site of the future deep rift basin. No precise dates are known for the beginning of sedimentation. Sedimentation ended with the eruption of the earliest rift basalts, which flowed over unconsolidated sands at about 1,109 Ma. Judging by the thinness of the sandstones and their fluvial nature, the onset of sedimentation probably did not much precede 1,109 Ma.

VOLCANIC ROCKS

One of the most striking features of the MRS is the great volume of subaerial flood basalts and related andesite and rhyolite that was erupted during rifting. The volcanic rocks accumulated to thicknesses of as much as 20 km in the deepest parts of the rift (Behrendt and others, 1988; Cannon and others, 1989). Estimates of the volume of volcanic rocks preserved in the rift range from 1.3 to 1.5 million km^3 (Hutchinson and others, 1990; Cannon, 1992), and the original erupted volume is estimated to be at least 2 million km^3 (Cannon, 1992). These volumes are comparable to those of many Phanerozoic flood basalt provinces.

Despite some complications in the Mamainse Point sequence, most of the earliest MRS rocks have reversed magnetic polarity and are overlain by rocks of normal polarity. This change in polarity presumably at about 1,097 Ma is a critical time marker by which geographically separated volcanic sequences are correlated where precise radiometric ages are lacking. Figure 32 summarizes the stratigraphic correlation of the Keweenawan volcanic sequences in the Lake Superior region.

AGE

Based on U-Pb zircon dating of volcanic rocks at the base of the Osler Group and the related Logan diabase sills in Ontario, rift-related magmatism began about 1,108.8+4/–2 Ma during a period of reversed magnetic polarity (Davis and Sutcliffe, 1985). The magnetic polarity shifted from reverse to normal between 1,097.6±3.7 Ma and 1,096.2±1.8 Ma (Davis and Sutcliffe, 1985; Davis and Paces, 1990). At Mamainse Point, a second stratigraphically thin reverse-normal sequence presumably occurs within the lower reversed sequence. Klewin and Berg (1990) have shown that this is not a structural repetition of the section as originally postulated by Palmer (1970). The remaining volcanic section was erupted during the ensuing period of normal polarity. Rift-related magmatism ceased shortly after the intrusion of a normally polarized rhyolite porphyry dated at 1,086.5+1.3/–3.0 Ma (Palmer and Davis, 1987) on Michipicoten Island in Lake Superior. Although volcanism was active for about 23 m.y., the rate of accumulation of volcanic rocks, at least in the central graben, increased steadily from 1,109 Ma to about 1,095 Ma, and thereafter waned as the rift became a sediment-dominated depression. Most volcanic rocks accumulated during the 3–5 m.y. interval, shortly after 1,098 Ma (Davis and Paces, 1990). Intrusive activity also was focused during the 23 m.y. interval in which the volcanic rocks were extruded. The Duluth Complex, for example, was intruded in stages between 1,107 and 1,096 Ma (Miller and others, 1991).

STRATIGRAPHY

The oldest volcanic rocks of the MRS are exposed in several areas, including the base of the Osler Group and the Mamainse Point Formation in Ontario, at the base of the North Shore Volcanic Group in Minnesota, and in the Powder Mill Group in Michigan and Wisconsin (fig. 32). Locally the basal flow is pillowed, but otherwise all flows appear to have been subaerial. In the western part of the Lake Superior region, the basal MRS section generally conformably overlies thin early-rift quartzites. In the eastern Lake Superior region, the early quartzites are absent and the basal flows were deposited unconformably on Early Proterozoic and Archean rocks.

COMPOSITION

Keweenawan volcanic rocks range in composition from olivine tholeiite to rhyolite. Some of the oldest lavas have a distinctly different mineralogy and chemistry than younger Keweenawan basalts. These basal rocks are transitional to weakly alkaline olivine basalts characterized by low Al_2O_3 contents and clinopyroxene phenocrysts. Clinopyroxene phenocrysts are rare higher in the sequence. Other basal lavas that lack clinopyroxene phenocrysts have higher Al_2O_3 contents and may be slightly younger than the clinopyroxene-bearing basal lavas (Nicholson and others, 1991). Locally, basal lavas are picritic, as at Mamainse Point (Berg and Klewin, 1988) and in some of the early intrusions in the Lake Nipigon region (Sutcliffe, 1985).

By far the dominant rock type among Keweenawan volcanic rocks is high-Al olivine tholeiite (Al_2O_3=15–19 wt. percent) followed by lesser high-Fe tholeiite and rocks of intermediate and felsic composition (Green, 1982; Brannon, 1984; Paces, 1988). The olivine basalts commonly are ophitic in texture, and the dominant phenocryst is plagioclase. The most primitive Keweenawan basalts are geochemically similar to primitive mid-ocean ridge basalts. However, incompatible trace elements in most Keweenawan basalts are enriched compared to depleted or primitive mantle. Radiogenic isotope analyses (Sr, Nd, and Pb) of the main stage high-Al olivine tholeiites suggest that a likely source of the voluminous basalts is a mantle plume (Paces and Bell, 1989; Nicholson and Shirey, 1990).

Flanking basins contain as much as 10–25 percent rhyolite (for example, North Shore Volcanic Group), whereas the deep central grabens contain minor (<1 percent) rhyolite (for example, Portage Lake Volcanics). Chemical and isotopic analyses of rhyolites suggest that they were derived from variable mixtures of melts from several sources, including Archean crust, Early Proterozoic crust, and previously erupted Keweenawan basalts (Brannon, 1984; Dosso, 1984; Nicholson and Shirey, 1990).

INTRUSIVE ROCKS

Intrusive rocks associated with the Midcontinent rift include alkaline intrusions, diabase dikes and sills, and layered intrusions and associated felsic rocks (Weiblen, 1982).

ALKALINE INTRUSIONS

Intrusions of carbonatite-alkalic complexes are presently known only north of Lake Superior, along the Kapuskasing structural zone, a zone of upthrusted amphibolite- to granulite-grade rocks of the Superior province, and along a north-trending zone north of the Coldwell Alkalic Complex (Weiblen, 1982; Sage, 1987, 1988). Weiblen (1982) has summarized the important characteristics of the 10 nephelinitic/carbonatitic complexes in the southern part of the Kapuskasing zone. Carbonatitic complexes such as the Prairie Lake carbonatite (1,030 Ma; Bell and Blenkinsop, 1987) typically are roughly circular, less than 2 km in diameter, and composed of carbonatite (mostly calcite) around the margins and ijolite (nepheline-pyroxene rock) in the core of the intrusion (Sage, 1987).

Three major alkaline complexes in the vicinity of Lake Superior include the Coldwell (1,108–1,099 Ma; Heamen and Machado, 1987), Killala Lake, and the Firesand River Alkalic Complexes (1,060 Ma; Bell and Blenkinsop, 1987). The Coldwell Complex is the largest, with a diameter of 28 km, and is located west of the Kapuskasing zone (Weiblen, 1982). The complex consists of a series of three intrusions containing gabbro, nepheline syenite, and syenite as the major rock types (Mitchell and Platt, 1978; 1982). These bodies are roughly concentric; gabbro forms the rim and grades into syenite in the cores. Other alkalic complexes north of the Coldwell Alkalic Complex range in diameter from 1 to 5 km but contain the same rock types, with or without minor carbonatite (Weiblen, 1982).

DIKES AND SILLS

Dike and sill swarms related to Keweenawan rifting are preserved in regional groups around the margins of present-day exposures of rift rocks in the Lake Superior region, and extend as far as central Wisconsin. Major swarms around Lake Superior occur in the Baraga-Marquette area of Michigan, in the Mellen-Gogebic area of Wisconsin and Michigan, in Carlton County and near Duluth in Minnesota, on the west edge of the Duluth Complex at Babbitt and Ely-Moose Lake in Minnesota, and near Grand Portage and Pigeon Point in Ontario and Minnesota; others are the Logan sills in Ontario and the Pukaskwa swarm along northern Lake Superior (Green and others, 1987). The dike orientations are generally subparallel to the axis of the rift, but in some areas multiple generations of dikes crosscut one another.

Two broad groups of dikes have been distinguished by chemical composition and magnetic polarity. As a generalization, the oldest dikes are quartz tholeiites of reversed magnetic polarity, whereas the younger, normally polarized dikes are mostly olivine tholeiites (Weiblen, 1982; Green and others, 1987). However, a continuum of compositions is present among the dikes, and thus the correlation between dike composition and magnetic polarity is not rigid. Reversed-polarity dikes cut pre-Keweenawan rocks, but reversed-polarity dikes are not known to crosscut dikes of normal magnetic polarity. Normal-polarity dikes typically cut older Keweenawan rocks and are exposed closer to the rift than the older dike swarms (Green and others, 1987). Age relations are determined on the basis of crosscutting relations and magnetic polarity because ages have not yet been determined for most swarms (Davis and Sutcliffe, 1985).

Sill complexes are most commonly associated with the North Shore Volcanic Group in Minnesota and the Logan and Pigeon River intrusions to the north. The sills are as much as several hundred meters thick, whereas the dikes are several meters to tens of meters thick. Chemically, the sills fall into the two broad groups defined by the dike suites, quartz- and olivine-tholeiites. Four well-studied Logan sills are characterized by weak differentiation (Jones, 1984). Coarse-grained or locally granophyric zones commonly occur towards the top of the sills. At least some of the sills associated with the North Shore Volcanic Group are chemically more evolved than sills along the north shore of Lake Superior (Jerde, 1991). Also, Jerde (1991) has documented the uncommon occurrence of "reverse" zonation (that is, the most primitive material occurs in the center of the sill) in the Lester River sill near Duluth. Zircon from pegmatitic zones from two thick Logan sills of reversed magnetic polarity yield the oldest determined age for Keweenawan magmatism, 1,108.8+4/–2 m.y. (Davis and Sutcliffe, 1985).

LAYERED INTRUSIONS

The Duluth Complex in Minnesota, composed of a series of layered intrusions (Weiblen, 1982), is the largest of the intrusive rocks associated with the Midcontinent rift. The lower contact of the complex is bounded by Archean or Proterozoic rocks, and the upper contact is bounded by Keweenawan volcanic rocks, consistent with the hypothesis that the Duluth Complex was intruded beneath its volcanic edifice (Miller and Weiblen, 1990). It is not a typical layered intrusion, but is broadly characterized by early intrusions of anorthosite, which, in turn, are intruded by multiple intrusions of layered troctolite. (See Chalokwu and Grant, 1990, among others.) Both anorthosite and troctolite are cut by late ferrodiorite and granite-granophyre bodies, resulting in complicated stratigraphic relations. Ultramafic rocks are only a minor constituent of the complex, occurring as localized dunite bodies associated with troctolites and as several peridotite bodies with uncertain field relations. Near the base of the complex, low-grade copper-nickel sulfide mineralization took place. Other well-documented intrusive bodies associated with the Midcontinent rift include the Mellen Intrusive

Complex in Wisconsin (Olmsted, 1969; Klewin, 1988) and the Crystal Lake Gabbro in Ontario (Geul, 1970).

Late felsic intrusive rocks are only now beginning to receive study. It is difficult to uniquely tie felsic intrusions to associated mafic intrusions of the Duluth Complex or the Mellen Complex. The distinction between hypabyssal felsic bodies and felsic flows is blurred as the intrusive rocks give way to flows at the top of the section.

INTERFLOW SEDIMENTARY ROCKS

About 3–10 percent of the volcanic sequence is interflow sedimentary rocks. In the main native copper district of the Keweenaw Peninsula, 22 laterally extensive layers of sandstone and conglomerate were used as marker beds by the miners. Typically these units are red, immature, clastic rocks ranging from boulder conglomerate to mudstone. The clasts are derived in large part from the local volcanic rocks. Structures such as graded bedding and crossbedding, imbrication, and ripple marks are locally well developed (Merk and Jirsa, 1982). Most interflow sediments were deposited as alluvial fans or in shallow lakes (Morey and Ojakangas, 1982).

POSTVOLCANIC SEDIMENTARY ROCKS

Lying on the volcanic rocks is a great thickness of fluvial clastic rocks and lesser lacustrine sedimentary rocks. The contact between sedimentary and volcanic rocks is generally conformable and commonly gradational, with the change characterized by more common and increasingly thick interflow sedimentary units upward in the volcanic section. The sedimentary rocks are most extensively exposed in northern Michigan and northern Wisconsin where the classic stratigraphic section has been defined and well studied. The following description is based mostly on summaries by Daniels (1982) and Dickas (1986).

The sedimentary succession is as much as 6 km thick in land exposures and is thicker beneath the lake as shown on seismic records. The lower part of the section consists of the Oronto Group. At the base of the group the Copper Harbor Conglomerate is largely coarse fluvial conglomerate near the base and grades upward into sandstone and pebbly sandstone. Units are uniformly reddish and thoroughly oxidized. Basalt flows are interlayered in the basal 1,000 m in some areas, indicating a transition from an environment of rifting and volcanism near the base to a more quiescent setting for upper parts of the unit. Clasts of all sizes are largely volcanogenic detritus, indicating derivation from highlands of only slightly older Keweenawan volcanic rocks. The Copper Harbor Conglomerate becomes finer grained toward the basin axis, an observation which, together with paleocurrent and other sedimentologic data, indicates deposition mostly on alluvial fans shed north from a prominent highland, which bounded the rift valley on the south.

The Nonesuch Shale conformably overlies the Copper Harbor Conglomerate; it is black to gray shale, siltstone, and fine sandstone, typically 100–200 m thick. A detailed internal stratigraphy, developed as a result of exploration and mining of extensive copper deposits near the base of the formation, can be traced over a very large region in northern

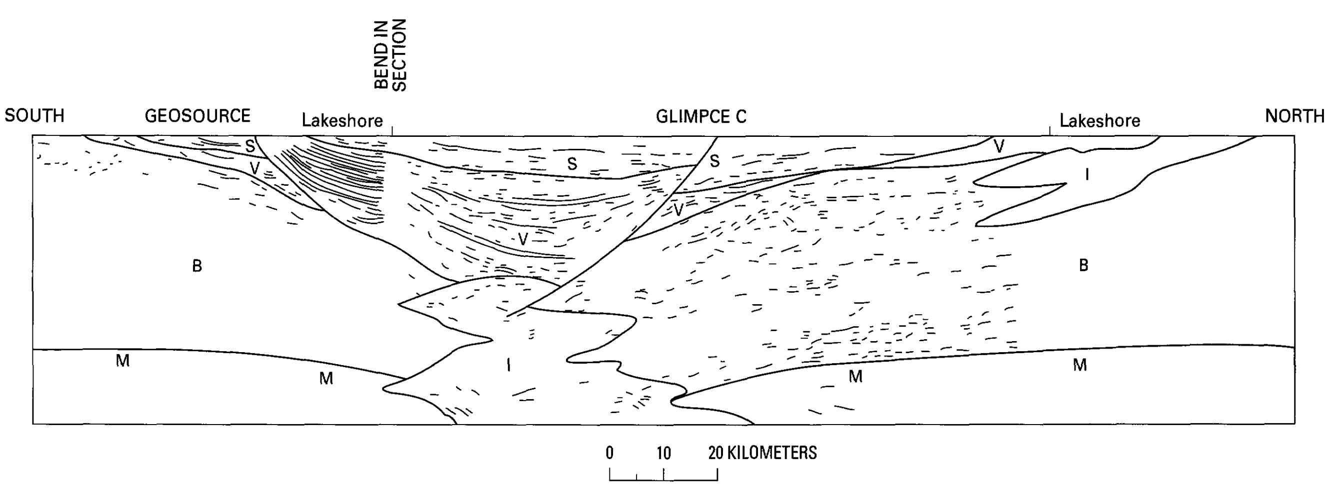


Figure 33. Crustal cross section of the Midcontinent rift. See figure 31 for location. S, rift sediments; V, rift volcanics; I, rift intrusions; B, basement of Archean and Early Proterozoic rocks; M, mantle. Form lines are strong seismic reflectors from GLIMPCE line C (Cannon and others, 1989) recorded to 20 s two-way travel time, and Geosource line south of Lake Superior (Hinze and others, 1990) processed to 5 s. Depth of reflectors is calculated for an average velocity of 6 km/s. No vertical exaggeration.

Michigan and Wisconsin, indicating deposition in a quiescent basin, probably a deep anoxic lake.

The overlying Freda Sandstone is mostly red to brown lithic sandstone, which generally becomes finer grained upward in the section. Volcanic fragments dominate the lithic clasts, indicating that much of the detritus was still being derived from Keweenawan volcanic rocks. Deposition was most likely a result of a variety of fluvial processes. Paleocurrent directions are generally toward or parallel to the rift axis, indicating that highlands of volcanic rocks existed south of the basin.

The youngest sedimentary rocks are the Jacobsville Sandstone in Michigan and the Bayfield Group in Wisconsin, taken as one unit, and the Fond du Lac and Hinckley Sandstones in Minnesota, seen as another. The two units are probably correlative but are nowhere in contact on land. Both units consist of fluvial reddish to buff, and locally white, sandstone and minor siltstone, shale, and conglomerate. They are mineralogically and texturally more mature than clastic rocks of the older Oronto Group. Detritus was derived from both Keweenawan volcanic and sedimentary rocks as well as from pre-Keweenawan rocks.

STRUCTURAL HISTORY

Recent seismic-reflection surveys (Behrendt and others, 1988; Cannon and others, 1989; Chandler and others, 1989; Hinze and others, 1990) reveal the internal structure of the rift in detail. Figure 33 illustrates a typical cross section of the rift based on a combination of seismic data and the exposed geology in northern Michigan and Minnesota. Subsidence occurred in response to crustal thinning in a complex manner, including development of an asymmetric graben, flexing of the crust, and ductile stretching of the lower crust. The result is deep central basins beneath which the prerift crust is nearly or totally separated. Accumulations of flood basalts attained 20 km thicknesses along the axis of the rift. Postvolcanic and postrift clastic rocks, nearly 10 km thick in places, were deposited in broad flexural basins centered on the rift graben. Basins are believed to have formed by flexure of a thin elastic crust during decay of the thermal anomaly caused by rifting and related magmatic activity (Nyquist and Wang, 1988).

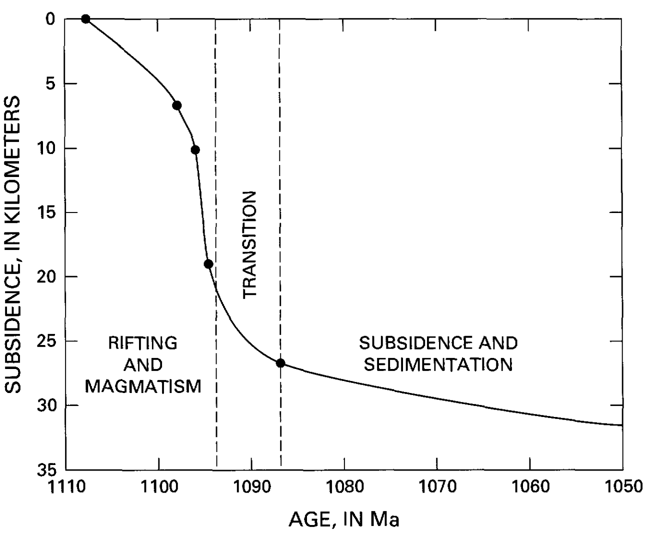


Figure 34. Subsidence curve for the axial region of the Midcontinent rift beneath Lake Superior. Solid circle, stratigraphic horizon for which precise U-Pb radiometric age has been determined. Right end of curve is less precisely constrained by age of syn-sedimentary faulting.

High precision U/Pb zircon radiometric ages (Davis and Sutcliffe, 1985; Palmer and Davis, 1987; Davis and Paces, 1990; Miller and others, 1991) provide excellent time resolution of events in the rift development. Together with seismic-reflection profiles, these ages permit the construction of subsidence curves such as figure 34 for the western part of Lake Superior. This curve illustrates the rapid rate but short duration of extension, rift subsidence, and accompanying volcanism, all of which persisted for a maximum of 15 m.y. Subsidence caused by crustal extension began at about 1,109 Ma and continued until about 1,094 Ma. The rate of subsidence accelerated throughout this period and the eruption rate of basalts was sufficient to more or less match subsidence, so that the central depression was always nearly filled with volcanic rocks. Extension, subsidence, and volcanism waned during the ensuing 10 m.y., during which sedimentation of coarse clastic rocks (Copper Harbor Conglomerate) dominated the filling of a broad basin which was evolving from an extensional to a flexural style of subsidence. The last eruptions produced relatively thin sequences of basalt flows and a few rhyolitic volcanic centers which were erupted until about 1,086 Ma. Broad subsidence and sedimentation lasted at least an additional 30 m.y., and probably substantially longer. During that time, the Nonesuch Shale was deposited in a lake of regional extent, and was followed by a return to a prolonged period of fluvial red-bed sedimentation during which the Freda Sandstone, Bayfield Group, and Jacobsville Sandstone were deposited.

A period of compression produced major reverse faults and broad folds during the latter phases of sedimentation. This faulting, of uncertain cause, reactivated earlier normal faults and partly inverted the original central graben. The age of the faulting has been dated indirectly as roughly 1,060±20 Ma (Ruiz and others, 1984; Bornhorst and others, 1988; Cannon and others, 1990b). Deposition of the younger parts of the Jacobsville Sandstone outlasted this faulting, giving a maximum limit for the age of the youngest sedimentary rocks.

DYNAMICS OF RIFTING

The origin of the forces that produced the MRS has been debated for many years. Opinion remains split between those who favor a totally passive mode of rifting in which far-field forces, such as would be generated at distant plate boundaries, are the cause (Donaldson and Irving, 1972; Chase and Gilmer, 1973; McWilliams and Dunlop, 1978; Gordon and Hempton, 1986) and those who favor rifting in response to a nearby mantle plume (Hinze and others, 1972; Burke and Dewey, 1973; Green, 1983; Cannon and Hinze, 1992). We argue herein that evidence for the involvement of a mantle plume in both the magmatic and dynamic evolution of the MRS is strong.

The presence of a mantle plume beneath the rift has been suggested by both petrochemical and isotopic studies of volcanic rocks (Paces and Bell, 1989; Nicholson and Shirey, 1990; Klewin and Berg, 1991) and by geophysical studies (Hutchinson and others, 1990). Cannon and Hinze (1992) proposed that the rift developed in response to the arrival of the head of a new mantle plume at the base of the lithosphere.

The great volume of basalt and only moderate degree of crustal extension, both conclusively documented by seismic-reflection profiles, led Hutchinson and others (1990) to conclude that a plume, which they named the Keweenaw hotspot, was present during rifting. Anomalously hot asthenosphere, as in the head of a plume, is required to account for the great volume of magma that was generated in the region.

A plume head also provides the large supply of fertile, primitive asthenosphere that seems to be required as a source for many of the basalts and intrusive rocks, to explain their primitive magma compositions and isotopic character.

Together, these two lines of evidence strongly argue for the presence of a plume in the region of the MRS as it developed at about 1.1 Ga. But whether the rift formed as a dynamic consequence of plume-generated stress, or developed as a result of other far-field plate margin stresses and coincidentally crossed the plume is more speculative.

Cannon and Hinze (1992) have pointed out many similarities between the development of the MRS and the sequence of events predicted by models of plume-generated magmatism and rifting developed by White and McKenzie (1989) and Campbell and Griffiths (1990). The enormous amount of magma generated in only a few million years, the scale of the region affected by rifting and volcanism (1,000–2,000 km diameter), and the prolonged period of thermal subsidence and sedimentation after rifting all are expected consequences of the arrival of a new plume head at the base of the lithosphere. Furthermore, the precise correlation in both time and place between plume-generated magmatism and rifting in the MRS argues strongly for a common cause for both.

In the model of Campbell and Griffiths (1990), as a new mantle plume rises through the deep asthenosphere, it develops a large bulbous head that contains both primitive mantle and more evolved upper mantle, which becomes entrained in the head. As the head approaches the base of the lithosphere, it spreads laterally to eventually underlie an area as much as 2,000 km in diameter, and undergoes partial melting as a result of adiabatic decompression. The arrival of the head results in a brief (10–20 m.y.) episode of intense volcanism, after which the energy in the head has been dissipated and volcanism ends. The arrival of the plume head also produces a transient, dynamically supported uplift of perhaps as much as 1 km. The gravitational potential of this uplift, although not adequate to produce complete continental breakup, may be adequate to have a more local and temporary effect on plate interiors (Hill, 1991).

Thus, we propose that, because the predicted consequences of the arrival of a new mantle plume beneath the lithosphere so closely match the documented history of development of the MRS, both the magmatism and extension of the MRS, including its southern extent, resulted from a new mantle plume that rose beneath the Lake Superior region at about 1.1 Ga.

EARLY AND MIDDLE PROTEROZOIC UPLIFT AND THERMAL EVENTS

By Z.E. Peterman *and* P.K. Sims

A variety of isotopic ages record three Early and Middle Proterozoic tectonic and "hydrothermal" events in the eastern part of the Lake Superior region. These events include a 1,625-Ma isotopic disturbance of regional extent, a major uplift in northern Wisconsin coeval with the main episode of Keweenawan igneous activity at 1,120 to 1,090 Ma along the Midcontinent rift system, and slightly younger thrusting that produced a 35-km-thick cross section of the Midcontinent rift. A less well defined event involving differential uplift of the Wolf River batholith and other areas is recorded by Rb-Sr and K-Ar biotite ages between 1,400 and 1,300 Ma.

THE 1,625 MA EVENT

The 1,625 Ma event is recorded by the resetting of several isotopic systems including Rb-Sr whole-rock and mineral ages of Archean and Early Proterozoic rocks over

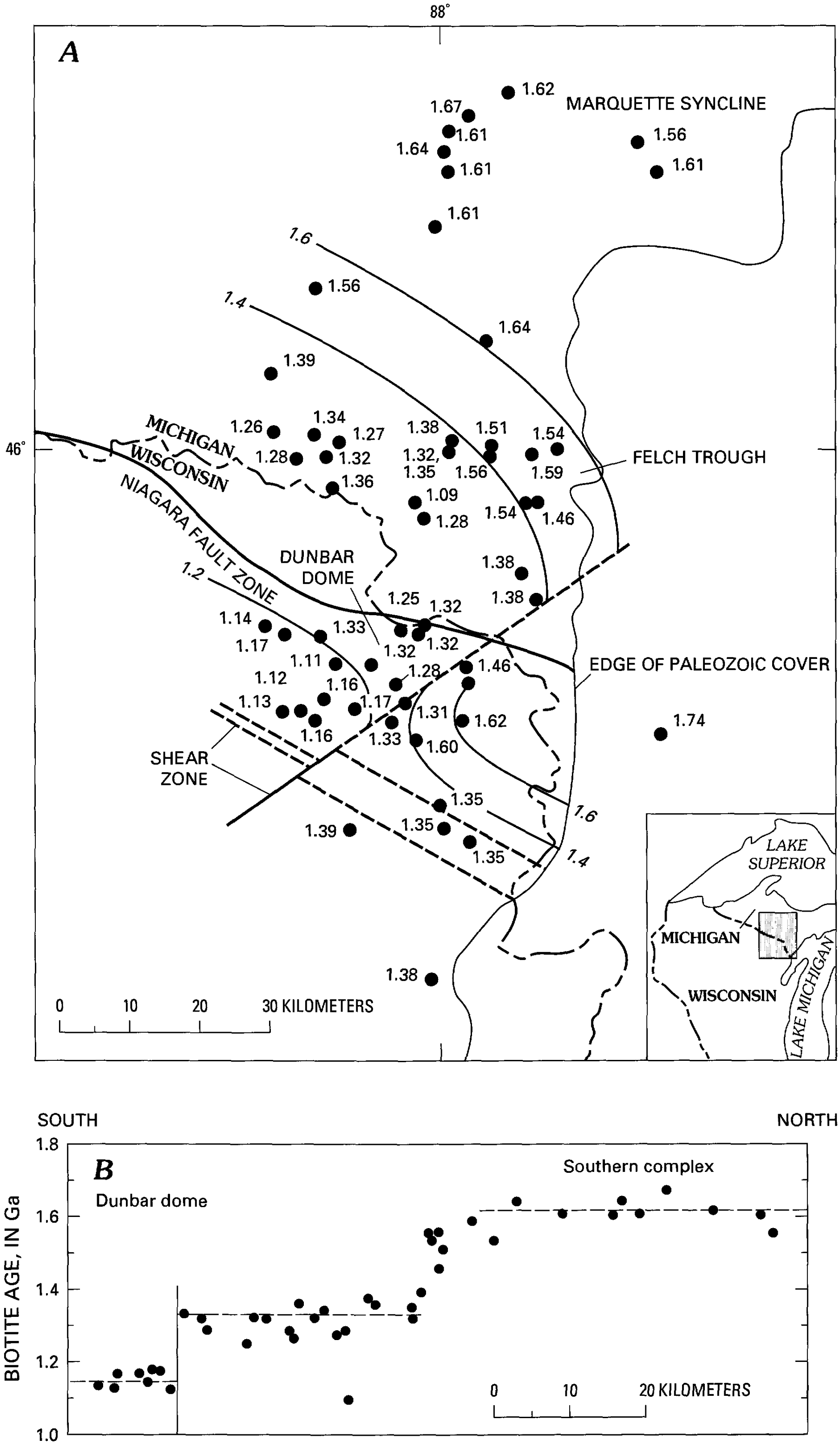


Figure 35. Rb-Sr biotite ages in transect from Marquette syncline, Upper Peninsula, Michigan, to northeastern Wisconsin. Modified from Peterman and others, 1985. *A*, Map showing ages relative to major geologic features. *B*, Profile of Rb-Sr ages projected to a N. 45° E. section from Marquette syncline to northeastern Wisconsin. Only ages at localities northwest of the northeast-trending shear zone are included. The step-function nature of the biotite age pattern is evident.

Table 2. Rb-Sr analyses of samples of the Michigamme Formation, Upper Peninsula, Michigan, and northeastern Wisconsin.

Sample No.	Type	Rb, ppm	Sr, ppm	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
206-85	Concretion	57.06	717.2	0.2305	0.71118
205A-85	Graywacke	40.88	197.5	0.6003	0.71955
205G-85	Graywacke	54.68	209.9	0.7556	0.72310
205H-85	Graywacke	57.39	148.5	1.122	0.73127
205C-85	Argillite	128.4	119.7	3.130	0.77863
205F-85	Argillite	140.7	103.3	3.979	0.79946
205I-85	Argillite	128.8	59.6	6.352	0.85328

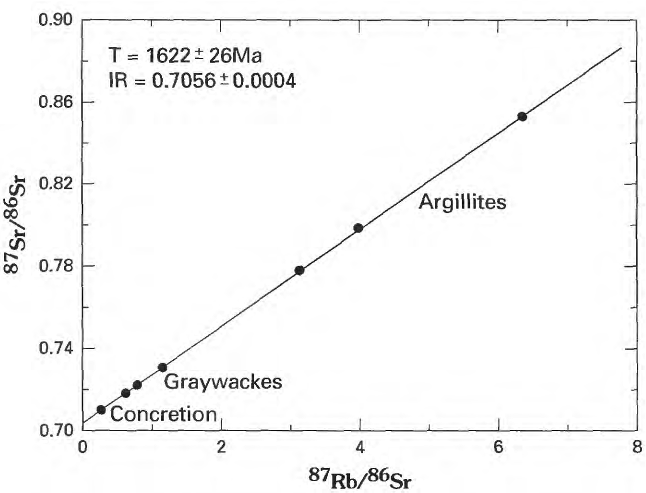


Figure 36. Rb-Sr whole-rock isochron plot for samples of argillite, graywacke, and carbonate concretion from Michigamme Formation, near Covington, Mich. The data indicate resetting of the Rb-Sr systems at 1,622±26 Ma.

much of the eastern part of Michigan's Upper Peninsula and northern Wisconsin (fig. 35). The weighted mean of 12 published whole-rock and mineral isochron ages is 1,625±25 Ma (Peterman and others, 1985, table 4). The mean was calculated using ISOPLOT (Ludwig, 1991) with the assumption that the ages compose a single population. An MSWD (mean square of weighted deviates) of 2.65 indicates only modest dispersion in excess of the errors assigned to each age. Exclusion of the two lowest ages results in a more precise mean of 1,635±17 Ma with an MSWD of 0.868.

To further document the 1,625 Ma event, we collected a suite of metagraywacke and argillite samples from the Early Proterozoic Michigamme Formation in the vicinity of Covington, Mich., west of the Marquette syncline at about long 88° W. (fig. 35*A*). At this locality, the Michigamme is tightly folded with east-striking fold axes dipping steeply to the south. The metamorphic grade is low (chlorite and sericite in the matrix), and primary structures and textures, including features of soft-sediment deformation, are well preserved.

Samples were collected from a dynamited road cut where graded graywacke beds, commonly between 0.045 and 0.1 m thick, are intercalated with shale beds less than 0.045 m thick. Three samples each of argillite and metagraywacke were selected for Rb-Sr analyses (table 2). The data give a precise isochron of 1,622±26 Ma (fig. 36), which is identical within error to the weighted mean age of 1,625±25 Ma for the regional data sets. Data for a sample of a carbonate-rich concretion plot on the isochron within error, but it is not included in the regression.

Bedded phosphorite from the base of the Michigamme in the Baraga basin (north of the Marquette syncline) has a concordant U-Pb age of 1,930 Ma (Marvin and others, 1988). Thus, the depositional age of the Michigamme near Covington is about 300 m.y. older than the reset age. Equivalent strata from the Watersmeet area, about 100 km to the west of Covington, which are at biotite and garnet grade, give a Rb-Sr isochron age of 1,820±50 Ma, which records metamorphism and deformation during the Penokean orogeny. In contrast, the Michigamme sample suite near Covington was not appreciably metamorphosed during the Penokean orogeny so the Rb-Sr whole-rock system was susceptible to subsequent disturbance at 1,625 Ma. In a study of Paleozoic graywackes in Australia, Graham and Korsch (1985) concluded that strontium-isotope equilibration in a graywacke-shale sequence occurred under temperatures and pressures as low as prehnite-pumpellyite to lowest greenschist facies conditions. Factors which facilitate isotopic homogenization include transformation of clays to chlorite and illite, albitization of plagioclase, and a high fluid flux (Graham and Korsch, 1989). Calcite is present in the Michigamme rocks as ubiquitous concretions and as granular carbonate in the matrix, and the position of the concretion sample on the graywacke-argillite isochron indicates that the calcite participated in the strontium isotopic exchange during the 1,622 Ma event.

Insofar as known, the 1,625 Ma event was amagmatic, and any sediments that may have been deposited either are unrecognized as being associated with the event or have been removed by erosion. In fact physical evidence of the event is meager. It is possible, however, that the mild deformation and metamorphism recorded in the 1,760 Ma rhyolite-granite terrane of southern Wisconsin occurred at this time. Van Schmus and Bickford (1981) have suggested that the 1,625 Ma event could be the result of stresses transmitted from a convergent plate margin far to the south. Possibly the disturbance is related to accretion of the Early Proterozoic Central Plains orogen (Sims and Peterman, 1986) to the North American craton.

THE 1,400–1,300 MA EVENT

Subsequent events, superimposed on the region perturbed at 1,625 Ma, mainly involved differential uplift. Evidence for one or more uplift events in the interval 1,400–1,300 Ma is provided by biotite ages of crystalline rocks in the Republic, Mich., area and in the 1,469 Ma Wolf

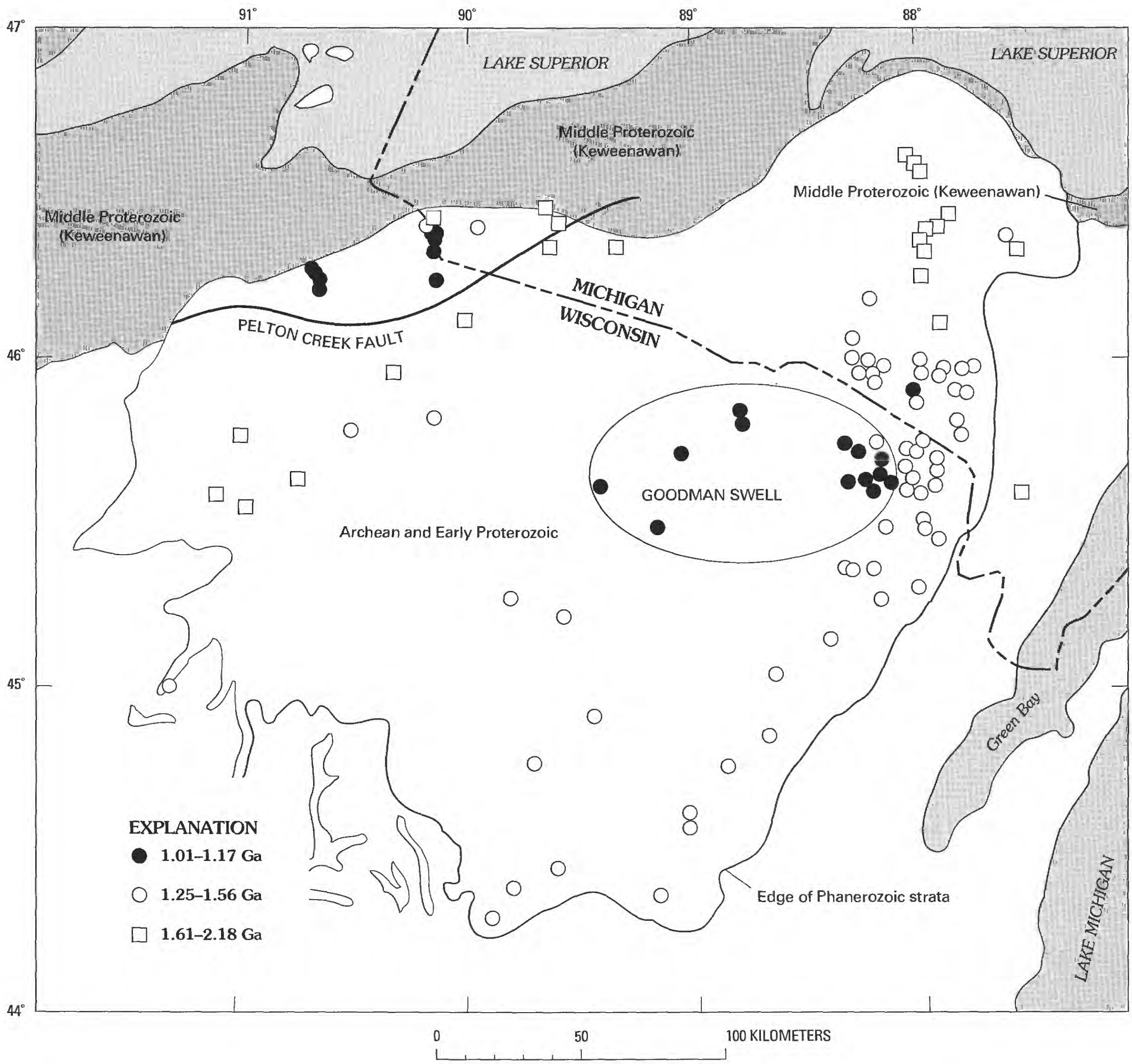


Figure 37. Rb-Sr biotite ages of Archean and Early Proterozoic rocks in northern Wisconsin and Upper Peninsula of Michigan. The elliptical area bounded by the 1.2 Ga contour is the Goodman swell.

River batholith in eastern Wisconsin. Rb-Sr biotite ages within the area extending from the Marquette syncline to northeastern Wisconsin (fig. 35*A*) define a systematic pattern that youngs southward in discrete age steps (Peterman and others, 1985). In an area extending from the Felch trough to the central part of the Dunbar dome, Rb-Sr and K-Ar biotite ages define an "age" plateau at 1,330±42 Ma (1,250 to 1,390 Ma) (fig. 35*B*).

Rb-Sr biotite ages of seven samples of the Wolf River batholith collected on a northeast-southwest axial traverse average 1,381±11 Ma (1,364 to 1,396 Ma), whereas four biotite ages in country rock near the contact average 1,439±16 Ma (fig. 30). The age difference and the existence of shear zones along the contact suggest differential uplift of the batholith about 100 m.y. after its emplacement (Peterman and Sims, 1988). Possibly, the uplift of the Wolf River batholith and the area in the vicinity of Republic are coeval, but the mean ages lack sufficient resolution to affirm or deny consanguinity.

THE 1,128±20 MA EVENT

A large area in northern Wisconsin underlain by Early Proterozoic crystalline rock was uplifted during Keweenawan rifting. The core of the uplifted area recorded by the 1.1 Ga biotite ages can be circumscribed by an ellipse (shown by a 1.2 Ga contour) having a west-trending major axis of 100–150 km and a minor axis of 50–100 km (fig. 37). This area has been termed the Goodman swell (Peterman and Sims, 1988). Biotite ages increase away from the swell in all directions, attaining values in excess of 1.6 Ga in distant parts of the Upper Peninsula of Michigan and northwestern Wisconsin.

The mean Rb-Sr age of biotites within the Goodman swell of 1,128±20 Ma overlaps the main pulse of volcanic activity within the Midcontinent rift system (1,109–1,090 Ma), suggesting a causal relation between rift activity and closure of the Rb-Sr biotite systems. Peterman and Sims (1988) concluded that the Rb-Sr age pattern within the Goodman swell resulted from rapid uplift of this area during development of the Midcontinent rift system. They postulated that the swell resulted from flexuring of the lithosphere in response to loading and deflection of the crust along the axis of the rift. The existence and location of the swell resulted from the southward concavity of the integrated axial rift load and the resultant flexural interactions (Peterman and Sims, 1988, fig. 5).

One additional area where biotite ages indicate Keweenawan uplift is the region immediately south of the Gogebic range (fig. 37). The rocks with reset ages are in the upper plate of a thrust fault system of crustal dimensions that was active shortly after rifting (Cannon and others, 1990a,b, 1993). The upper plate rocks were tilted northward during southward thrusting on listric faults; subsequent erosion exposed depths where the temperature was in excess of the closure temperature for biotite. The thrusting produced a 35-km-thick cross section of the Midcontinent rift. The age of faulting corresponds closely with compressional events (≈1,060 Ma) in the Middle Proterozoic Grenville province to the east (Cannon and others, 1993). A zone of lithospheric weakness along the newly formed rift may have become the locus for lithospheric shortening in response to stresses transmitted from the Grenville province. In the same way, the Goodman swell could have resulted from the same stresses inasmuch as it was uplifted essentially concurrently with the thrust faulting.

MAFIC DIKE SWARMS

By P.K. Sims

Major mafic dike swarms of two ages occur in the United States segment of the Lake Superior region. A northwest-trending dike swarm comparable in size to the other great dike swarms of the Superior province (Fahrig and West, 1986) is older than the Early Proterozoic Animikie Group, and smaller, less homogeneous dike swarms are Middle Proterozoic (Keweenawan) in age (Green and others, 1987). The pre-Animikie dikes, named the Kenora-Kabetogama (KK) dike swarm, transect Archean rocks in northern and western Minnesota (Southwick and Halls, 1987) and adjacent Ontario throughout an area that is at least 300 km wide and 300 km long. The Keweenawan dikes cut the Middle Proterozoic rocks of the Midcontinent rift system and adjacent Early Proterozoic and Archean rocks.

KENORA-KABETOGAMA DIKE SWARM

Mafic dikes that compose the KK swarm form a remarkably consistent and persistent northwest-trending set

Table 3. Mafic dike swarms of Middle Proterozoic (Keweenawan) age in Lake Superior region.

[nd, not determined; R, reversed magnetic polarity; N, normal magnetic polarity. Modified from Green and others, 1987]

Locality	Symbol (fig. 38)	Magnetic polarity	Trend
Baraga and Marquette Counties, Michigan.	BM	R, rare N	East-west.
Carlton County, Minnesota.	CC	R, few N	N. 30° E.
Grand Portage, Minnesota.	GP	R	N. 65°–90° E.
Logan, Minnesota	L	R	N. 65° E.
Pukaskwa, Ontario	P	R, few N	N. 50° W.
Duluth, Minnesota	D	N	North-south.
Pigeon River, Minnesota.	PR	N	N. 65° E., N. 40° W.
Babbitt, Minnesota	B	N	N. 50°–55° W.
Ely–Moose Lake, Minnesota.	EM	nd	N. 65°–70° E.
Mellen-Gogebic, Wisconsin.	MG	R and N	N dikes, N. 20° E. R dikes E. to NE.
Keweenaw Peninsula, Michigan.	KP	nd	NE. to ENE.
Central Wisconsin	CW	R and N	N. 60°–85° E.

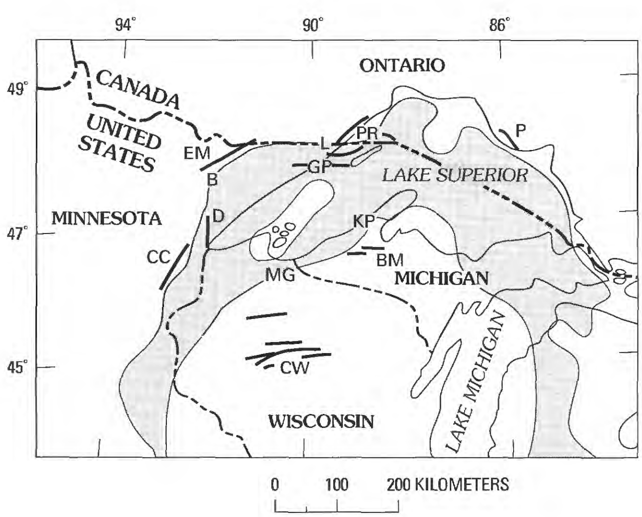
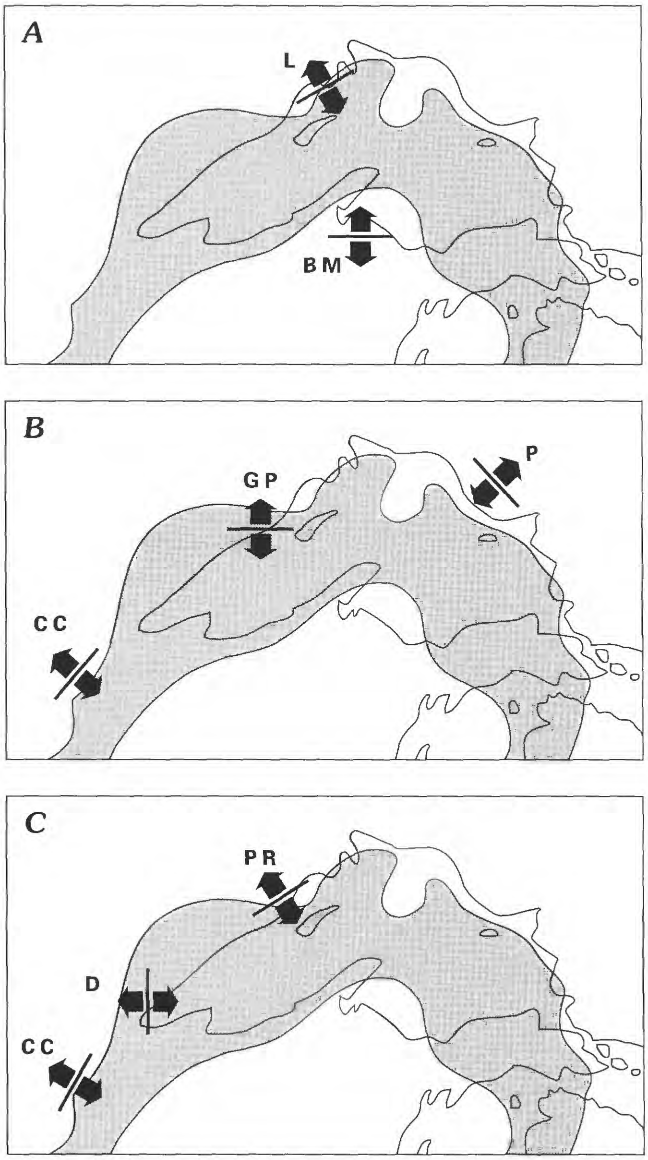


Figure 38. Mafic dike swarms related to Middle Proterozoic Midcontinent rift system in Lake Superior region. Patterned area, positive gravity anomaly associated with the rift. Modified from Green and others, 1987. B, Babbitt; BM, Baraga-Marquette; CC, Carlton County; CW, central Wisconsin; D, Duluth; EM, Ely–Moose Lake; GP, Grand Portage; KP, Keweenaw Peninsula; L, Logan; MG, Mellen-Gogebic; P, Pukaskwa; PR, Pigeon River.

Figure 39 (facing page). Sketch maps showing evolution of Midcontinent rift system as shown by the temporal sequence and structure of mafic dike swarms in Lake Superior region. Patterned area, positive gravity anomaly associated with the rift. Modified from Green and others, 1987. *A*, Initial opening of rift in central part of rift arc, shown by Logan (L) and Baraga-Marquette (BM) R swarms. *B*, Next rifting phase, shown by Pukaskwa (P), Grand Portage (GP), and Carlton County (CC) R dikes. *C*, Third phase of rifting, after magnetic polarity reversal at ≈1,100 Ma, producing Pigeon River (PR), Duluth (D), and Carlton County (CC) N dikes. Final phase, producing the Portage Lake Volcanics, not indicated because its feeder dikes are still covered. Large arrows, direction of rifting.



in northern Minnesota that fans somewhat to a more westerly trend in west-central and southwest Minnesota (fig. 47; also Chandler, this report). The dikes are sharply delineated on the shaded-relief aeromagnetic maps of Minnesota (Chandler, 1991a). The dikes are basaltic in composition and have two distinct compositions that appear to represent a continuum in magma composition (Southwick and Halls, 1987). One group has MgO and TiO_2 contents of about 7.6 and 0.9 percent, respectively, and the other group has MgO and TiO_2 contents of about 5.5 and 2.2 percent, respectively. The chemical compositions plot as a linear array that transects the boundary between the fields of high-Mg and high-Fe tholeiite on the Jensen diagram. The composition of the dikes is not that of "typical" continental basalt, despite their continental setting. The KK compositions are impoverished in Al_2O_3, relative to the continental field in the discrimination diagram of Pearce and others (1977), and plot mainly in the ocean-island field of the diagram. This composition may indicate that the mafic magma moved quickly to high levels in the granitoid crust, where it froze.

The high-Ti dike suite has a Rb-Sr isochron age of 2,120±67 Ma (Beck and Murthy, 1982), whereas the low-Ti suite has an inferred K-Ar age of about 2,240 Ma (Hanson and Malhotra, 1971).

Paleomagnetic data (Halls, 1986) indicate that the high-Ti and low-Ti dikes have pole positions that are statistically indistinguishable. These data imply that both dike groups were emplaced during an interval of time when the North American craton and the magnetic pole remained essentially fixed with respect to each other.

The KK dike swarm is spatially and temporally associated with the Penokean orogen (Southwick and Day, 1983). Southwick and Halls (1987) have suggested that the dikes may be related to peripheral uplift and extension on the cratonic side of the Proterozoic flexural foredeep (Animikie basin). The slightly radial pattern of dikes with respect to the tectonic front of the Penokean orogeny is consistent with this hypothesis.

MIDDLE PROTEROZOIC (KEWEENAWAN) DIKES

Mafic dike swarms associated with the Midcontinent rift system generally trend subparallel to the arcuate rift system and reflect complex extension related to opening of the rift. They are only locally well exposed, and the swarms have been given local, mainly geographic names (fig. 38). The magnetic polarity and trend of the dikes for the major swarms are listed in table 3.

Paleomagnetic studies have established the existence of a significant reversal in magnetic polarity during the Keweenawan interval: an older reversed (R) polarity and a younger normal (N) polarity. They also have established a migration of paleomagnetic poles along an apparent polar wander path going southwest to south as part of the western arm of the so-called "Logan loop" (Robertson and Fahrig, 1971). Thus, R dikes are considered to be older than N dikes, and can be related to R lavas; N dikes are inferred to be feeders for younger N lavas.

The Keweenawan dikes of both R and N polarities have compositions typical of continental tholeiites (Green and others, 1987). The oldest activity (as indicated by pole positions on the R part of the polar wander track) is recorded in the Logan and Baraga-Marquette swarms (fig. 39*A*), which suggest that rifting began near the center of what is now Lake Superior. This phase was succeeded by dike emplacement in the Pukaskwa swarm, the Grand Portage swarm, and the Carlton County swarm (fig. 39*B*).

After the polarity reversal, rifting resumed with formation of the N polarity Pigeon River, Duluth, and Carlton County dikes (fig. 39*C*). The youngest lava sequence (Portage Lake Volcanics) was erupted from fissures that are still deeply buried, although the N dikes in the Mellen-Gogebic swarm may be related to this magmatism. The age and the paleomagnetic relations of the central Wisconsin and Ely–Moose Lake swarms are not known. The dikes in central Wisconsin mainly occupy pre-Keweenawan faults.

TECTONIC EVOLUTION

By P.K. Sims

The Precambrian rocks in the Lake Superior region record six major coherent episodes of crust generation. The nominal ages of these events are: 3.55–2.6 Ga, 2.75–2.6 Ga, 1.9–1.83 Ga, ≈1.76 Ga, 1.5–1.45 Ga, and ≈1.1 Ga. The three older events and the ≈1.1 Ga event were episodes in which primitive material was added to the crust from the mantle or juvenile lower crust. The ≈1.76 Ga and 1.5–1.45 Ga events were episodes of anorogenic magmatism that followed the Penokean orogeny, and involved partial melting of crustal rocks.

The structural makeup of the Superior province has been interpreted, mainly from studies in Canada, as resulting from subduction-driven accretion of Archean crustal elements, most of which range in age from 3.1 to 2.6 Ga (Card, 1990, and references therein). Accretion generally progressed from north to south. A high-grade gneiss terrane in the north is succeeded southward by a broad region of alternating volcanoplutonic (greenstone-granite) and metasedimentary subprovinces of mainly Late Archean age, which produce a distinctly linear pattern. The southernmost terrane, the Minnesota River Valley subprovince (gneiss terrane), has been interpreted as having been accreted at about 2.69 Ga along the Great Lakes tectonic zone (Sims and Day, 1992), a major suture earlier suggested as the terminal collision in the Superior province (Hoffman, 1989).

The Late Archean subprovinces in the United States, from north to south, the Wabigoon, Quetico, and Wawa subprovinces (pl. 1), are of mantle and juvenile crustal origin with little evidence of inheritance of components older than ≈3.1 Ga. The volcanic rocks (2.75–2.70 Ga) represent oceanic, island arc, and continental arc volcanism; volcanism was accompanied by mafic to trondhjemitic plutonism. The metasedimentary belt (Quetico subprovince) comprises variably metamorphosed volcanogenic turbidites having detrital zircons ranging in age from about 3.0 Ga to 2.7 Ga. These rocks are intruded by abundant granitoid rocks. The metasedimentary rocks were deposited after major volcanism and possibly represent accretionary prisms, although turbidites of the Wawa and Quetico subprovinces in northern Minnesota appear to grade laterally into one another (Southwick, this report). The three juvenile subprovinces in the United States have collectively been called "greenstone-granite terrane" in previous reports on the Lake Superior region, to distinguish them clearly from the Archean gneiss terrane, now referred to as the Minnesota River Valley subprovince of the Superior province.

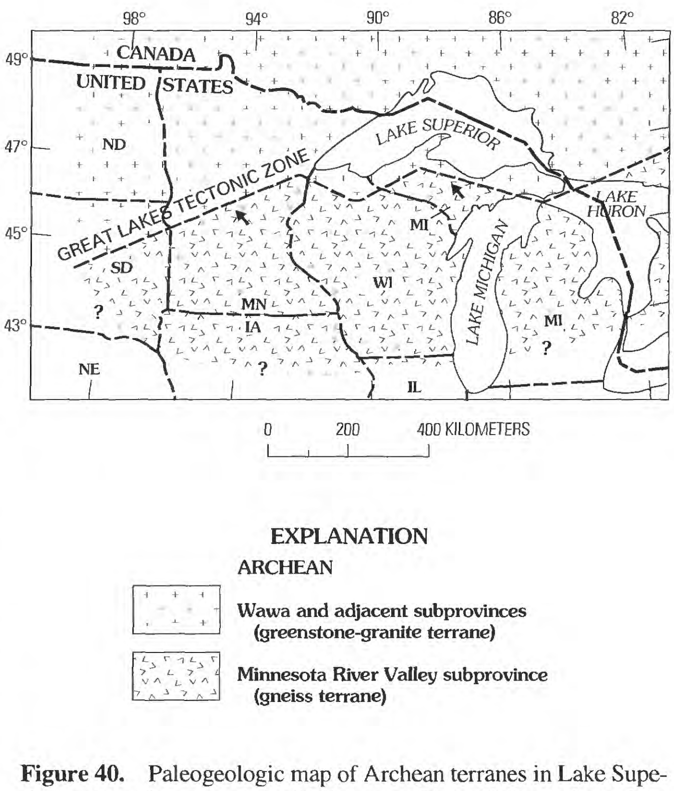


Figure 40. Paleogeologic map of Archean terranes in Lake Superior region after accretion of Minnesota River Valley subprovince (gneiss terrane) to south margin of Wawa subprovince (Superior province) along Great Lakes tectonic zone. The rocks are restored to their approximate prerift (Keweenawan) position; their geographic positions are based on present-day orientations and relative positions inasmuch as their position on the Earth in pre-Keweenawan time is not known. Modified from Sims, 1976, 1987. The gneiss terrane represents a remnant of a moderately large sialic protocontinent. Arrows, inferred direction of tectonic transport during suturing; extent of terranes queried where uncertain.

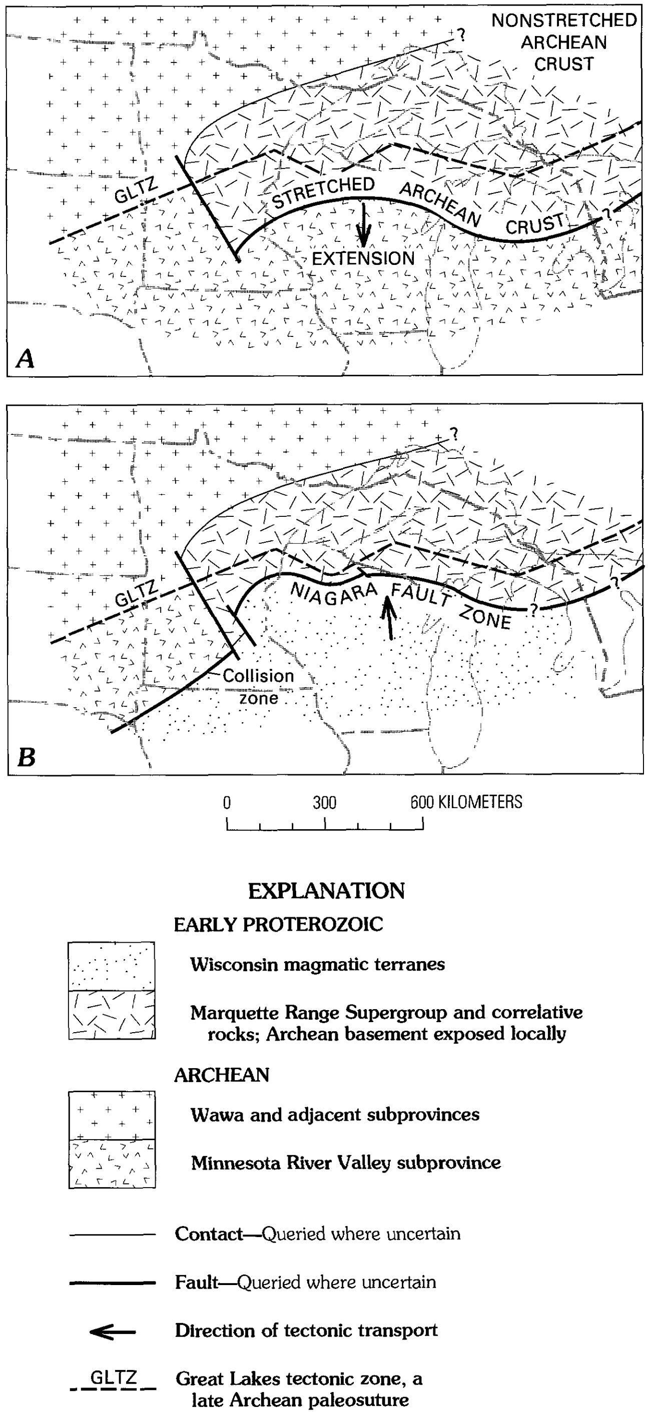


Figure 41. Paleogeologic maps showing successive stages of development of Early Proterozoic Penokean orogen, Lake Superior region. Modified from Sims and others (1987). Same reconstruction as figure 40. *A*, Early Proterozoic epicratonic sequence (1.9–2.1 Ga), showing tensional regime and inferred outline of depositional basin. Sediments and volcanics of the Marquette Range Supergroup and equivalents in Minnesota were deposited on stretched Archean crust in the Lake Superior region, and somewhat older (2.5–2.2 Ga) sediments of the Huronian Supergroup (of Canada) were deposited in the Lake Huron area. Western extent of epicratonic sequence is uncertain. *B*, Arc-continent collision (1.85 Ga) along the Early Proterozoic Niagara fault zone. Collision followed rifting of the continental margin, development of a south-dipping subduction zone with attendant island arc (or back-arc basin) volcanic and granitoid rocks, and closing of the ocean basin between the arc terrane and the continental margin. Collision produced foredeep deposits and a foreland fold-and-thrust belt on the continental margin.

The Archean Minnesota River Vàlley subprovince (gneiss terrane) (3.55–2.6 Ma), represents a remnant of a large sialic protocontinent (fig. 40) inferred to have extended from the Lake Superior region eastward into the Lake Huron region (Sims and others, 1980). Its original southern and western extents are not known. This terrane records a long and protracted tectonic and magmatic history. The gneisses in the Benson, Montevideo, and Morton blocks in southeastern Minnesota are about 3,500 Ma, as is gneiss in the Watersmeet dome, northern Michigan (Sims and others, 1984). Gneiss in the Jeffers block (pl. 1), to the south, however, is about 2,700 Ma. Also, except for the gneiss in the Watersmeet dome, all gneisses that have been dated in northern Michigan are Late Archean in age; some of these rocks are supracrustal, as for example the Dickinson Group (Sims, 1992). Existing isotopic age data suggest that Early Archean crust is limited to the northern part of the exposed Archean gneiss terrane, and that much of the terrane is composed of Late Archean gneiss. Late tectonic granite of ≈2,600 Ma intrudes the older gneisses. The gneiss terrane was partly reactivated during the Penokean orogeny, as indicated by gneiss domes and block uplifts of that age in northern Michigan and east-central Minnesota (pl. 1).

Data from the Marquette, Mich., area (Sims, 1991a) indicate that at this locality suturing along the Great Lakes tectonic zone resulted from oblique, northwest-oriented dextral thrust-shear that produced southward subduction of the juvenile Wawa subprovince rocks and (or) northward obduction of the Archean Minnesota River Valley subprovince (gneiss terrane). The inferred direction of tectonic transport during suturing is shown by solid arrows in figure 40. The collision can account for the development of the numerous steep, dextral faults and associated sinistral faults formed by transpression (Hudleston and others, 1988) in the segment of the Superior province north of the GLTZ, many of which host valuable gold deposits (Sims and Day, 1992).

After stabilization and strengthening of the crust (2.5 Ga), a southward-thickening, asymmetrical sequence of sedimentary and bimodal volcanic rocks (2.1–1.9 Ga) accumulated on the rifted craton margin in the Lake Superior region (fig. 41), and a somewhat older (2.5–2.2 Ga) sequence was deposited in the Lake Huron area, Canada (fig. 41*A*). These sequences were deposited on stretched Archean crust and constitute a passive margin sequence.

Following breakup of the North American continent—and the development of oceanic crust, most of which has been destroyed—complex island arcs (and (or) back-arc basins) were formed to the south (present-day coordinates). Arc volcanic and granitoid rocks (Pembine-Wausau terrane of Wisconsin magmatic terranes) were formed during southward subduction of rocks on the continental margin, and near the suture, collisional granitoid rocks were formed (Sims and others, 1992). Collision of this arc complex with the continental margin along the Niagara fault zone at 1.85 Ga (fig. 41*B*) produced a pronounced foreland fold-and-thrust belt on the continental margin, telescoping Archean basement rocks and their Early Proterozoic supracrustal cover across a width of 85 km. Erosion of the northward-verging thrust blocks containing arc rocks of the Pembine-Wausau terrane produced northward-migrating foredeep deposits that were deposited on and were structurally intercalated over a limited period of time with the older cratonic sedimentary and volcanic rocks of the continental margin (Barovich and others, 1989).

About 10 m.y. after terminal continent-arc collision along the Niagara fault zone, subduction was reversed, resulting in accretion of the Marshfield terrane (of Wisconsin magmatic terranes) to the Pembine-Wausau terrane along the Eau Pleine shear zone (pl. 1). Supracrustal volcanic (≈1.86 Ga) and sedimentary rocks of the Marshfield terrane overlie an Archean (3.0–2.7 Ga) gneissic basement. The Archean rocks could represent an exotic terrane. Post-tectonic alkali-feldspar granite bodies (1,835 Ma) intruded the suture zone as stitching plutons.

At ≈1.76 Ga, rifting on northeast-trending axes was accompanied by deposition of largely subaerial, anorogenic rhyolite and cogenetic epizonal granite in southern Wisconsin and adjacent areas. The basement is presumably Early Proterozoic and Archean rocks of the Marshfield terrane (pl. 1). This anorogenic activity is an older component of anorogenic igneous activity in the Midwest that progressively youngs southward (Sims and others, 1987; Sims, 1990f). Approximately concurrent with the anorogenic magmatism, calc-alkaline granodiorite to granite plutons were emplaced into the Pembine-Wausau terrane (Sims, 1992). The 1.76 Ga magmatism is approximately coeval with development of the Central Plains orogen (Sims and Peterman, 1986; Sims and others, 1987; Sims, 1990d; Sims, Peterman, and others, 1991) to the south, and may be a consequence of accretion of this terrane to the North American continent.

Quartz-arenitic red-bed sequences were formed both before and after the ≈1.76 Ga anorogenic magmatism, as determined by paleomagnetism (Chandler and Morey, 1992) and indirect dating. The present consensus is that quartzite bodies were formed both during later stages of the Penokean orogeny and in post-1.76 Ga time. The older quartzite sequences are more intensely deformed than the younger ones, and may in part be allochthonous (LaBerge and others, 1991). The quartzite sequences probably were deposited as braided fluvial deposits on a sandy coastal plain that subsequently was submerged during a shallow marine transgression (Dott, 1983). Southwick and Mossler (1984) demonstrated that the Sioux Quartzite in southwestern Minnesota was deposited in fault-bounded, intracratonic basins.

In the interval 1.5–1.45 Ga, the Wolf River batholith and associated anorogenic intrusions in central Wisconsin were emplaced as part of the continent-wide transcontinental anorogenic province. These plutonic rocks are thought to be deeper seated (mesozonal), coeval equivalents of the ≈1.48 Ma rhyolite exposed in the St. Francois Mountains and vicinity in Missouri (Sims and others, 1987).

The youngest crust-forming event in the Lake Superior region was the magmatism (≈1.1 Ga) and subsequent sedimentation within the Midcontinent rift system. No consensus exists concerning the origin of the rift system, but the tremendous amount of mafic magma that was erupted and its chemical and isotopic composition favor a hot-spot origin (Cannon and Nicholson, this volume).

GRAVITY AND MAGNETIC STUDIES CONDUCTED RECENTLY

By Val W. Chandler[2]

Because Pleistocene glacial drift covers most of the Precambrian bedrock, geophysical surveys have been an integral part of geologic studies and mineral exploration in the Lake Superior region. The greatest success has been with gravity and magnetic methods; the data are reasonably inexpensive to acquire, and both methods are highly sensitive to many of the lithologic variations in the Precambrian bedrock. Gravity and magnetic data have been used extensively in geologic mapping of the Precambrian bedrock surface and in many cases have provided the only available means of interpreting geology in the third dimension. Several reports have reviewed earlier work in the Lake Superior region and have described the general relationships of anomaly signatures to geology (Chandler, 1993; Hinze and Wold,

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1982; Klasner and others, 1985; Sims, 1972a, b). This report will focus on work from about 1985 to the present, a time in which gravity and magnetic synthesis of the Lake Superior region progressed significantly.

Much of the progress in gravity and magnetic studies of the Lake Superior region can be attributed to new gravity and magnetic data sets. Some of these data sets involve acquisition of new gravity and magnetic data, whereas others are wholly compilations of preexisting data. The higher resolution allowed by new surveying has enhanced detailed geologic mapping in many areas, and the uniform, broad scale perspective allowed by the larger compilations has been helpful to regional scale mapping and correlations. The data sets from these recent studies are digitized, allowing a broad spectrum of computer applications to process, display, and interpret the data.

HIGH-RESOLUTION AEROMAGNETIC SURVEYS

Two recent aeromagnetic surveys have contributed greatly to our knowledge of the geophysics and geology of the Lake Superior region; one survey covered the State of Minnesota and the other covered Lake Superior. Both surveys produced high-resolution data that superseded earlier, regional scale coverage.

A recently completed program of high-resolution aeromagnetic surveying (fig. 42), conducted by the Minnesota Geological Survey (MGS) between 1979 and 1991, has significantly enhanced studies of the Precambrian geology of Minnesota. Additional high-resolution data were contributed by the U.S. Geological Survey, USX Corporation, and the Geological Survey of Canada. The U.S. Geological Survey flying was conducted as part of the Conterminous United States Mineral Assessment Program (CUSMAP) for the International Falls and Roseau 1°×2° sheets (fig. 42). The new data, most of which were acquired with flightline spacings of 380–536 m and mean terrain clearances of 91–213 m, represent more than 560,000 line kilometers of flying; this supersedes earlier regional scale coverage by the U.S. Geological Survey (Zietz and Kirby, 1970). The high-resolution data have been merged and gridded at an interval of 213.36 m (Chandler, 1991a,b), and a shaded relief image of these data is shown in figure 43. Geologic interpretation of the high-resolution aeromagnetic data has been supplemented by completion of the regional (2–3 km average spacing) gravity coverage for the entire State (Chandler and Schaap, 1991).

Aeromagnetic coverage over Lake Superior has been greatly improved by a recent high-resolution aeromagnetic survey of Lake Superior (Teskey and others, 1991), which was done as a contribution to the Great Lakes International Multidisciplinary Program on Crustal Evolution

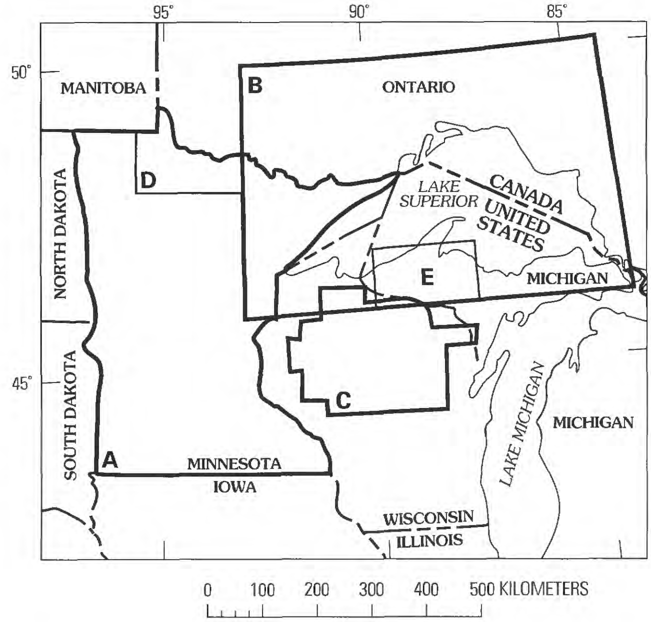

Figure 42. Index map showing areas of new geophysical data acquisition and compilation in Lake Superior region. Subareas outlined by heavy lines represent aeromagnetic data shown in figures as follows: A, Minnesota high-resolution aeromagnetic data (fig. 43); B, GLIMPCE aeromagnetic survey and regional compilation (fig. 44); C, recompilation of aeromagnetic data from Wisconsin (King, 1990) (fig. 45). Subareas outlined by light lines represent USGS-CUSMAP surveys and compilations as follows: D, Roseau and International Falls CUSMAP, new acquisition of high-resolution aeromagnetic data and some gravity data; E, Iron River CUSMAP, compilation of previously existing data.

(GLIMPCE). The surveying was conducted by the Geological Survey of Canada using a line spacing of 1.9 km and an altitude of 305 m above lake surface. The new aeromagnetic data supersede earlier, regional scale coverage of Lake Superior (Wold and Ostenso, 1966; Hinze and others, 1966; O'Hara, 1982; Hinze and Wold, 1982). The Lake Superior data were combined with data surrounding the lake by the Geological Survey of Canada, the U.S. Geological Survey, and the Minnesota Geological Survey (fig. 42) and gridded at a 400 m interval. A shaded relief image of the gridded data is shown in figure 44. A companion gravity data set and grid are presently being compiled from existing gravity data in the region (Kucks and others, 1989).

OTHER RECENT COMPILATIONS OF GRAVITY AND AEROMAGNETIC DATA

Recently compiled gravity and aeromagnetic data sets have been produced in the Lake Superior region by the U.S. Geological Survey as a part of its Conterminous United States Mineral Assessment Program (CUSMAP). For the

Figure 43. Shaded relief map of high-resolution aeromagnetic data in Minnesota. Illumination from the northwest at 45° inclination. GLTZ, Great Lakes tectonic zone.

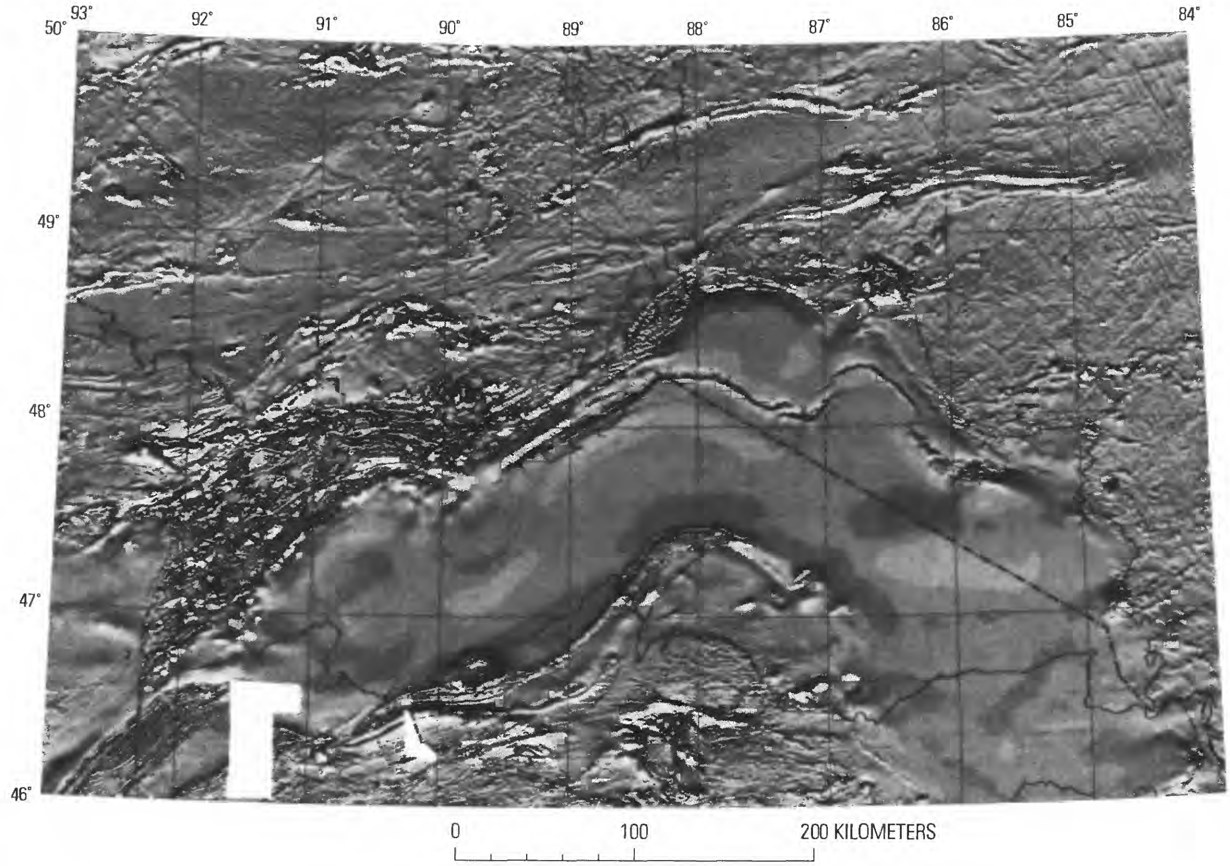

Figure 44. Shaded relief map of aeromagnetic data from the GLIMPCE survey, Lake Superior and adjacent areas. Illumination from south at 40° inclination. Blank area, no data. (From Teskey and others, 1991.)

Iron River quadrangle of Michigan and Wisconsin (fig. 42, subarea E), the gravity (Klasner and Jones, 1989) and aeromagnetic (King, 1987) compilations were mainly from previously existing data; for the International Falls and Roseau quadrangles in Minnesota, the gravity (Horton and Chandler, 1988a, b) and aeromagnetic (Bracken and Godson, 1987; Bracken and others, 1989a, b) compilations involved some acquisition of new data. New data also were obtained for the Hibbing quadrangle (Bracken and Godson, 1988). The high-resolution aeromagnetic data of the International Falls and Roseau sheets are incorporated into the Minnesota data shown in figure 43, and the aeromagnetic data of the Iron River sheet are incorporated into the Lake Superior data shown in figure 44.

The aeromagnetic data from central and northern Wisconsin (Zietz and others, 1977; Karl, 1986) were recompiled into 200 m grid and presented as a shaded relief map by King (1990). Because outcrop and drill hole control are meager over much of central and northern Wisconsin, these data have played an important role in recent interpretations of the Precambrian geology of northern and central Wisconsin (Mudrey and others, 1987; Sims and others, 1989; Sims, 1992). A shaded relief image of the Wisconsin aeromagnetic data is shown in figure 45.

Some large regional compilations of gravity and magnetic data have also assisted geologic studies in the Lake Superior region. These include those for the central United States (Hildenbrand and others, 1983; Hildenbrand, 1989) and for North America (Committee for the Gravity Anomaly Map of North America, 1988; Committee for the Magnetic Anomaly Map of North America, 1988). These data provided a basis for revision of the Precambrian basement geology of the continental interior (Sims, 1990d; Sims, Peterman, and others, 1991), and help place the geology of the Lake Superior region in a broader context of continental structure and evolution (Hildenbrand, 1985; Sims and others, 1987).

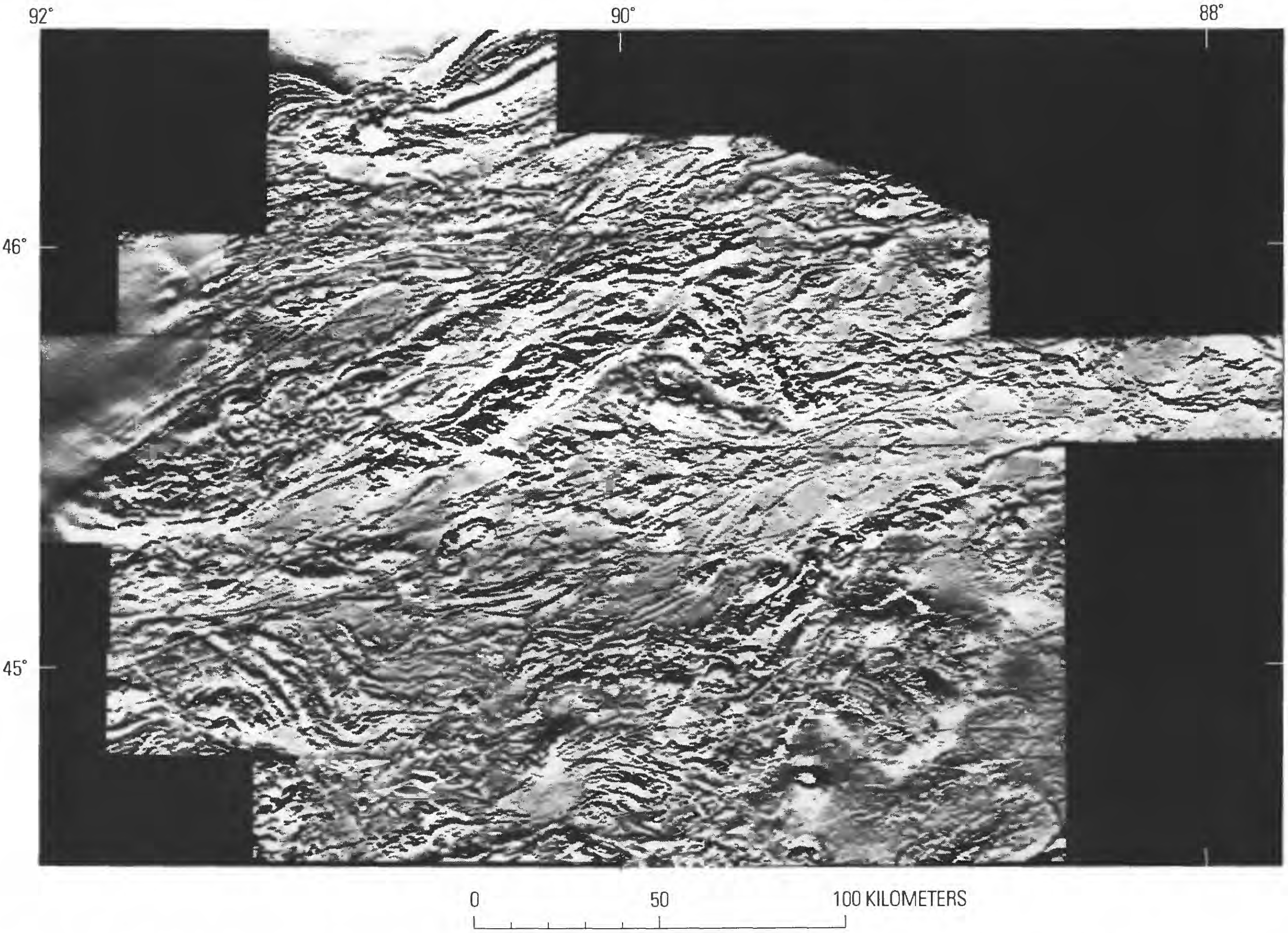

Figure 45. Shaded relief map of aeromagnetic data from northern Wisconsin (King, 1990). Illumination from south at 89.5° inclination.

INTERPRETATION OF GRAVITY AND MAGNETIC ANOMALIES

New gravity and magnetic data sets have been important adjuncts to recent geologic studies in the Lake Superior region. The applications have included simple qualitative use of anomaly maps to delineate broad-scale lithologic units and structures, as well as quantitative analysis to estimate physical characteristics of anomaly sources. In spite of the ambiguities inherent to gravity and magnetic interpretations, the methods have proven to be powerful tools when used in conjunction with existing geologic and rock property data. The discussion that follows regarding magnetic signatures refers to the shaded relief aeromagnetic maps in figures 43, 44, and 45. For discussion of gravity signatures, a gravity anomaly map of the Lake Superior region is presented in figure 46.

MINNESOTA RIVER VALLEY SUBPROVINCE (ARCHEAN GNEISS TERRANE)

High-grade metamorphic rocks of the Minnesota River Valley subprovince (Archean gneiss terrane) of southwestern Minnesota are characterized by a generally positive gravity expression that is locally interrupted by lows of a few tens of milligals (fig. 46), and by a complex magnetic signature that includes large-amplitude (500–1,000 nanoteslas) positive magnetic anomalies (fig. 43). The gneiss terrane is divided into three distinct blocks by east- to northeast-striking magnetic lineaments that most likely represent shear zones (Chandler and Southwick, 1990; Schaap, 1989; Chandler and Schaap, 1992), and is bounded on the north by the Great Lakes tectonic zone (pl. 1; fig. 43). Gravity and magnetic modeling (Schaap, 1989) indicates that the proposed shear zones, as well as the overall internal structures of the gneissic blocks, generally dip at moderate

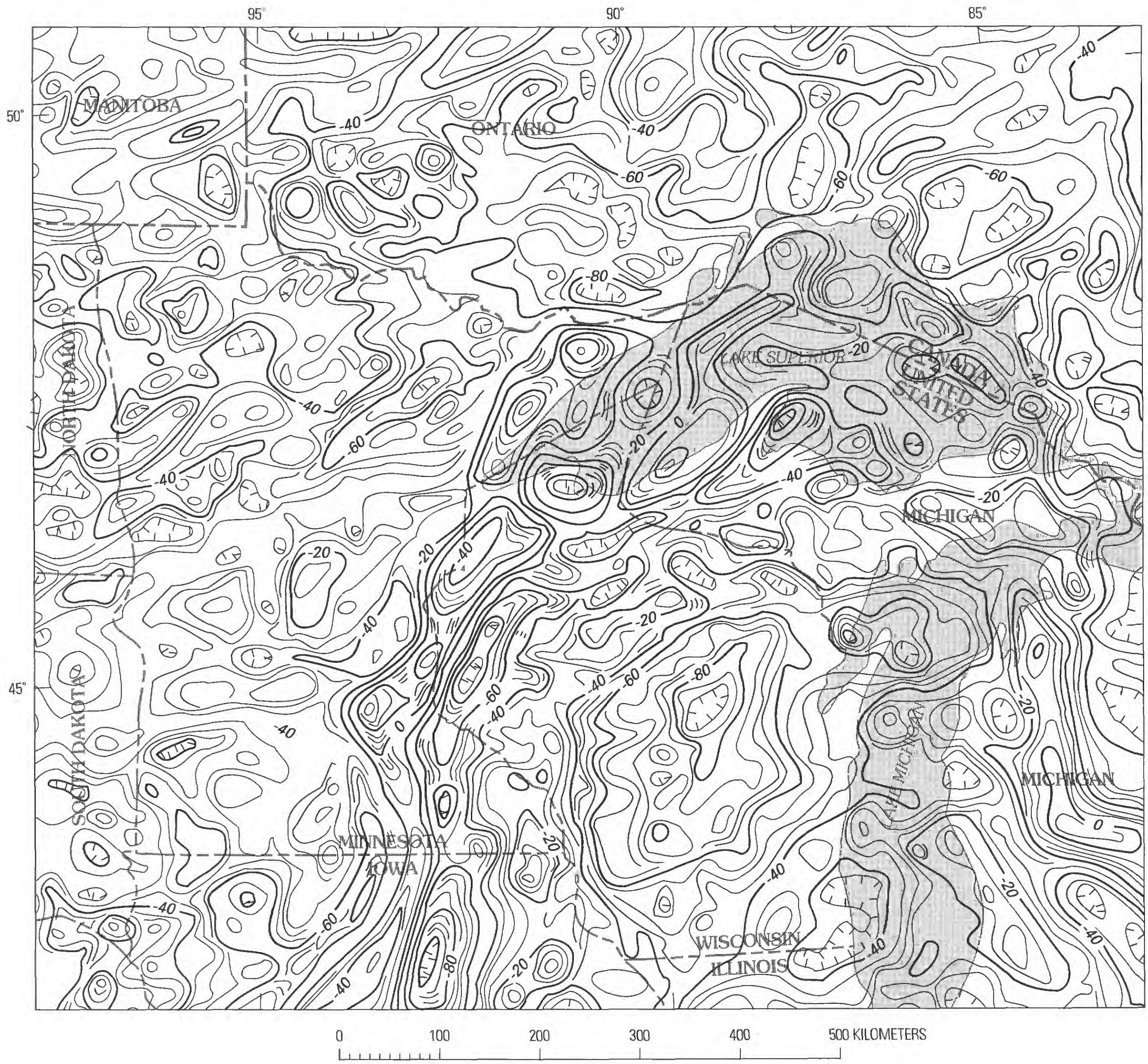

Figure 46. Bouguer gravity anomaly map of Lake Superior region. Contour interval 5 and 20 mGals; hachures indicate area of closed low (from Committee for the Gravity Anomaly Map of North America, 1988).

angles to the north, and they are therefore concordant with a north-dipping seismic reflector attributed by Gibbs and others (1984) and Smithson and others (1986) as being the Great Lakes tectonic zone. The southern terminus of the gneiss terrane is not well defined, but sparse geologic control and gravity and magnetic signatures imply that extreme southwestern Minnesota may be underlain by a different terrane that consists chiefly of Early Proterozoic supracrustal and intrusive rocks (Schaap, 1989). These rocks are in turn overlain by basins of Early Proterozoic Sioux Quartzite (Southwick and others, 1986), which produce patches of subdued magnetic signature in extreme southwestern Minnesota (fig. 43).

WAWA AND ADJACENT SUBPROVINCES (ARCHEAN GREENSTONE-GRANITE TERRANE)

The gravity and magnetic signatures of the Wawa and adjacent subprovinces (Archean greenstone-granite terrane) are best developed in northern Minnesota. Here the magnetic signatures are characterized by northeast strikes and include curvilinear highs with very large amplitudes (>1,000 nanoteslas) over greenstone-hosted iron-formations and broad, irregular highs with moderate amplitudes (100–500 nanoteslas) over granitic bodies (fig. 43). Intervening belts of subdued magnetic signature are generally associated with metasedimentary rocks. The gravity signature of the greenstone-granite terrane in northern Minnesota consists of northeast-striking lows and highs with amplitudes of a few tens of milligals (fig. 46), which correspond to low-density granitic and high-density metavolcanic belts, respectively.

Strong physical property contrasts and steeply dipping contacts in the Archean greenstone-granite terrane produce gravity and magnetic anomaly signatures that are ideally suited for geologic mapping. All recent geologic mapping in the greenstone-granite terrane in Minnesota (Day and others, 1990, 1991; Jirsa, 1990; Jirsa and Boerboom, 1990; Jirsa and others, 1991; Jirsa and others, 1994) and Michigan (Cannon, 1986; Sims, 1992) has relied extensively on gravity and aeromagnetic data. Although this work primarily involves qualitative use of gravity and magnetic maps, it has also included quantitative interpretation through processing of gridded data and modeling along profiles in Minnesota (Case and others, 1990; Jirsa and others, 1991) and Michigan (Klasner and Jones, 1989).

EARLY PROTEROZOIC PENOKEAN OROGEN

Gravity and magnetic studies have been an integral part of the recent reinterpretation of the Penokean orogen as resulting from continent-continent collision (Southwick and Morey, 1991; Sims and others, 1989). Qualitative analysis of gravity and magnetic maps has been particularly useful in the recognition and mapping of faults and major geologic units, most of which are very poorly exposed. Quantitative interpretation of gravity and magnetic data through grid processing and modeling along profiles has also made significant contributions.

EAST-CENTRAL MINNESOTA

In east-central Minnesota, Southwick and others (1988) and Southwick and Morey (1991) have interpreted a prominent belt of east- to northeast-striking magnetic highs (500–>1,000 nanoteslas) to be part of a northward-verging fold-and-thrust belt which, on the basis of existing drill hole and outcrop data, includes several allochthonous terranes of Early Proterozoic supracrustal rocks (pl. 1; fig. 43). Areas of subdued magnetic signature to the north of this belt correspond to nonmagnetic rocks in the Animikie basin and associated outliers, which are interpreted to have been deposited in a tectonic foredeep along the front of the fold-and-thrust belt (Southwick and Morey, 1991). Inverse modeling of aeromagnetic data from east-central Minnesota using Werner deconvolution (Ferderer, 1988) indicates that many anomaly sources in the fold-and-thrust belt dip steeply northward to moderately southward, and that nonmagnetic strata in the Animikie basin could be as thick as 6 km near the front of the fold-and-thrust belt. Combined analysis of gravity and magnetic anomalies through Poissons theorem implies that the iron-formation–bearing Cuyuna terranes of the fold-and- thrust belt (Southwick and Morey, 1991) extend unconformably beneath relatively less deformed strata of the Animikie basin (Carlson, 1985; Chandler and Malek, 1991). This extension correlates with broad, east- to northeast-striking highs in the northern part of the Animikie basin (fig. 43; see also fig. 17) that reflect sources that are buried by 3–5 km of nonmagnetic Animikie Group strata (Ferderer, 1988).

NORTHERN MICHIGAN AND WISCONSIN

Recent compilations of gravity (Klasner and Jones, 1989) and aeromagnetic (King, 1987) data have been useful supplements to geologic mapping in northern Michigan and adjacent Wisconsin (Cannon, 1986; Sims, 1992). In this area, Early Proterozoic supracrustal rocks generally form open to troughlike basins that flank domal uplifts of Archean basement (pl. 1), and iron-formations and metavolcanic rocks in the supracrustal rocks produce prominent (>1,000 nanoteslas) magnetic highs that outline the uplifts (pl. 1; figs. 44, 45). These encircling magnetic highs are locally enhanced at metamorphic nodes, where the supracrustal rocks are metamorphosed to amphibolite grade (King, 1987). The troughlike basins of supracrustal rocks correspond to residual gravity highs of 2–15 milligals, and gravity modeling indicates that they are typically 2 to 5 km thick (Klasner and Sims, 1984; Klasner and others, 1985; Klasner and Jones, 1989).

The aeromagnetic data from Wisconsin (fig. 45) delineate many pertinent features within the poorly exposed magmatic terranes of the Penokean orogen (King, 1990). South of the Niagara fault zone, a 50-km-wide belt of strongly positive (500–1,000 nanoteslas) magnetic signature (pl. 1; fig. 45) characterizes mafic metavolcanic rocks of the Pembine-Wausau terrane (Sims and others, 1989). Ovoid areas of subdued signature within this belt delineate plutons of Penokean granitoid rocks. Some of these granitoid rocks correspond to 15–30 milligal gravity lows (fig. 46). A 25- to 50-km-wide zone of moderate amplitude (100–500 nanoteslas) magnetic signature to the south of the dominantly mafic volcanic rocks corresponds to metavolcanic rocks that have a significant felsic component. The Marshfield terrane (pl. 1), to the south, which consists of highly deformed Archean basement and Early Proterozoic supracrustal and plutonic rocks (Sims and others, 1989; Sims, 1992), corresponds to moderate-amplitude (generally 100–300 nanoteslas) magnetic signature that includes arcuate, northwest-striking anomalies in the southwestern part of the area of figure 45. Several major fault and shear zones within the Penokean magmatic terranes are defined by aeromagnetic lineaments; these include the Jump River shear zone, the Niagara fault zone, and the Eau Pleine shear zone (pl. 1; fig. 45; Sims and others, 1989). A gravity model that supported a seismic-reflection interpretation of a GLIMPCE line down the axis of Lake Michigan (Cannon and others, 1991) indicates that the Niagara fault and Eau Pleine shear zone dip steeply south and north, respectively, and both may penetrate the entire crust.

MIDDLE PROTEROZOIC WOLF RIVER BATHOLITH

The granitic rocks of the 1,470 Ma Wolf River batholith (pl. 1) correspond to a very subdued magnetic signature in the southeastern part of the area of figure 45. This area of subdued signature appears to be slightly below the anomaly base level of surrounding terranes, implying that some of the granitic rocks may be reversely polarized (King, 1990). Anorthositic rocks within the batholith and outlying plutons of granitic and syenitic rocks to the east (pl. 1) correspond locally to strongly positive (500–1,000 nanoteslas) magnetic signatures (King, 1990). Recent gravity model studies by Allen and Hinze (1992) indicate that the batholith is chiefly responsible for the regional negative gravity anomaly over central Wisconsin (fig. 46). In their model the batholith extends to depths of about 10 km and extends laterally almost 100 km to the northwest and southeast of its outcrop beneath a 2- to 4-km-thick cover of older rocks.

MIDDLE PROTEROZOIC MIDCONTINENT RIFT SYSTEM

Rocks associated with the Middle Proterozoic (1,100 Ma) Midcontinent rift system produce the most prominent gravity and magnetic signatures in the Lake Superior region (figs. 43, 44, and 46). Gravity and magnetic anomalies associated with the rift commonly exceed 50 milligals and 1,000 nanoteslas amplitudes, respectively. The strongly varying anomaly character reflects contrasts between mafic flows and intrusions, most of which contain a strong remanent magnetization, and low-density, nonmagnetic sedimentary rocks of the late-rift sequence. Gravity and magnetic data have been extensively used for geologic mapping and for model studies to investigate deep rift structure. Gravity and magnetic modeling has been an important supplement to seismic-reflection interpretations of the rift as determined by GLIMPCE (Cannon and others, 1989) and other projects (Hinze and others, 1992; Gibb and others, 1994).

WESTERN ARM

In eastern and southeastern Minnesota, strongly positive gravity and magnetic signatures indicate uplifted lavas of the Saint Croix horst, and areas of subdued magnetic signature that flank and locally occur on top of the horst correspond to basins of rift-related sedimentary rocks (pl. 1; figs. 43, 46). The northwest-striking segment of the rift in southeastern Minnesota corresponds to the Belle Plaine fault, where the smooth magnetic signature (fig. 43) reflects a deep fill of rift-related sedimentary rocks. South of the Belle Plaine fault the rift swings again to a northeast strike (pl. 2; fig. 46), forming the Iowa horst and associated basins.

Gravity and magnetic modeling of the western arm of the rift by McSwiggen and others (1987) and Chandler and others (1989) indicates that mafic rocks are largely restricted to the St. Croix horst (see fig. 31), where they may extend to depths of 10–20 km, implying that the horst originated as a central graben. These studies also indicated that sedimentary basins flanking the rift are 3–5 km thick near the horst margins, whereas the basins on top of the horst are 2–4 km thick. Modeling by Chandler and others (1989), which was constrained by seismic-reflection data, indicates that the reverse faults bounding the horst dip nearly vertical to moderately inward, and that some may have originated as growth faults during extrusion of the lavas. Magnetic modeling by Chandler and others (1989) indicates that most of the lava sequences defined by the seismic-reflection data are normally polarized, although some lavas deep in the succession of northwestern Wisconsin and central Iowa may be reversely polarized. Nyquist and Wang (1988) used gravity modeling in conjunction with seismic interpretation and flexural modeling to investigate the deep crustal structure beneath the western arm of the rift in Minnesota and Wisconsin. In the lower crust beneath the rift axis, they hypothesized a large mafic pillow, which was emplaced during rift magmatism and, upon cooling, may have been responsible for postvolcanic subsidence of the rift.

The models derived by McSwiggen and others (1987) and Chandler and others (1989) imply a marked asymmetry for segments of the rift with regard to off-axis placement of the deep mafic roots and postulated growth faults. Similar asymmetry has been interpreted for the rift segments to the south in Kansas (Woelk and Hinze, 1991) and to the north in Lake Superior (Cannon and others, 1989), and the general geometry has some similarities to the opposing half-graben model for continental rift propagation (Bosworth, 1985; Rosendahl, 1987). In such a scenario the northeast-striking segments of the St. Croix horst would represent half-graben basins and the Belle Plaine fault would represent a major accommodation zone.

DULUTH COMPLEX

In northeastern Minnesota the mafic intrusive rocks of the Duluth Complex (pl. 1) correspond to a strongly positive gravity signature (fig. 46) and large (>1,000 nanoteslas) positive and negative magnetic signatures (fig. 43). The complicated magnetic signatures associated with the Duluth Complex reflect a complicated intrusive history and the effects of locally strong remanent magnetization (Chandler, 1990). Derivative-enhanced gravity and aeromagnetic data have been useful in delineating intrusive units in the complex, including the poorly exposed interior (Chandler, 1990) and the partially exposed southeast margin (Miller and others, 1993). Gravity and magnetic modeling and analysis of derivative-enhanced aeromagnetic data along the northwest margin of the complex suggest that deposition of copper-nickel sulfides along the basal contact zone may have been enhanced by northwest-striking faults in the wall rock and by low dips of the basal contact over sulfur-bearing Animikie strata (Chandler and Ferderer, 1989). Gravity modeling in the same study indicates that the basal contact along the southwest margin of the complex dips steeply and probably is faulted. Gravity modeling of granophyre thicknesses has been used to help constrain differentiation models of mafic intrusions along the southeast margin of the complex (Miller and others, 1990).

LAKE SUPERIOR BASIN

Gravity and magnetic data have been important adjuncts to seismic-reflection profiling in Lake Superior. Seismic-reflection studies (Behrendt and others, 1988; Cannon and others, 1989; Hinze and others, 1990; McGinnis and Mudrey, 1991) indicate that as much as 30 km of volcanic and sedimentary rocks resides in asymmetric to symmetric grabens beneath western, central, and eastern Lake Superior, and that the bounding growth faults, which include the Douglas, Keweenaw, Isle Royale, and Michipicoten Island faults (pl. 1, fig. 31; Sims, 1992), were later reactivated as reverse faults with up to 2–5 km displacements. Gravity and magnetic modeling (Hinze and others, 1989, 1990; Mariano and Hinze, 1994a; Teskey and others, 1991; Teskey and Thomas, 1994; Thomas and Teskey, 1994) generally confirms this rift geometry, and indicates that a substantial part of the lava sequence beneath the central and eastern parts of the lake has reversed magnetic polarity. The traces of the bounding faults beneath the lake, which were originally based on low-resolution aeromagnetic and gravity data (Hinze and Wold, 1982), have been considerably refined by the new high-resolution aeromagnetic data of Lake Superior (fig. 44; Teskey and others, 1991; Mariano and Hinze, 1994b).

Seismic-reflection profiling verifies that two prominent gravity lows in western Lake Superior (fig. 46) represent thinning of the mafic volcanic sequence over rises in the prerift basement (Cannon and others, 1989; Mudrey and others, 1989; Coakley and others, 1992; Allen and others, 1994; Sexton and Henson, 1994), which was originally postulated by White (1966) in southwestern Lake Superior on the basis of geologic, aeromagnetic, and gravity data. Allen and others (1992) and Allen and Hinze (1992) isolated a regional (approx. 650 km wide) gravity minimum surrounding Lake Superior, and related it to isostatic compensation of a regional increase in surface elevation around the lake, which has been informally named the Lake Superior swell (Dutch, 1981b). Allen and others have postulated that the regional topography and negative gravity anomaly may reflect either a thickened crust or a depleted mantle, either of which could have been imprinted by a mantle plume during active rifting.

MAFIC DIKES AND SMALL PLUTONS

High-resolution aeromagnetic data in the Lake Superior region effectively delineate minor intrusions such as diabase dikes. These dikes typically produce linear magnetic anomalies with amplitudes of a few tens to a few hundreds of nanoteslas and are easily detectable on aeromagnetic maps. Although individually small, these intrusions commonly occur in regionally extensive swarms, and they therefore represent major tectonomagmatic events whose importance has generally been underestimated because of poor exposure. Figure 47, a map summarizing mafic dike swarms and plugs in the Lake Superior region, has been largely prepared from aeromagnetic data. Dike swarms shown in the Canadian segment of figure 47 will not be discussed here; they have been discussed recently by Fahrig and West (1986). The Early Proterozoic Matachewan dike swarm, which crosses the northeast corner of the area of figure 47, has been recently discussed by Bates and Halls (1991) and West and Ernst (1991).

The 2,120 Ma (Beck and Murthy, 1982) Kenora-Kabetogama dike swarm is expressed by a series of linear, northwest-striking anomalies that transect the magnetic signatures of the Wawa and adjacent subprovinces (Archean greenstone-granite terrane) and the Minnesota River Valley subprovince (gneiss terrane) of Minnesota (figs. 43, 47). The dikes are expressed most prominently in the greenstone-granite terrane in northwestern Minnesota (fig. 43). Based on the aeromagnetic data and exposures along its east margin, the swarm is at least 300 km long by 300 km wide and may contain thousands of individual dikes. Where exposed the dikes are typically 20–100 m thick, but preliminary model studies indicate that some of the dikes could be as thick as 200–300 m near the middle of the swarm (Southwick and Chandler, 1989). The aeromagnetic data (fig. 43) indicate that the dikes fan from trends of about 340° along their east margin to about 300° along the southwest margin, and the dikes converge towards the Penokean orogen, where they are truncated by the fold-and-thrust belt and are locally covered by foreland basin strata. Most of the dike anomalies in figure 43 are positive, but a few weakly negative anomalies near the west margin of the swarm may reflect a reversed paleomagnetic component that was observed in some exposed dikes along the east margin of the swarm (Halls, 1986). The Kenora-Kabetogama dikes may have been emplaced in response to an early rifting stage of the Penokean orogen (Southwick and Day, 1983), or in response to peripheral foreland bulging initiated by thrust loading of the crust in the

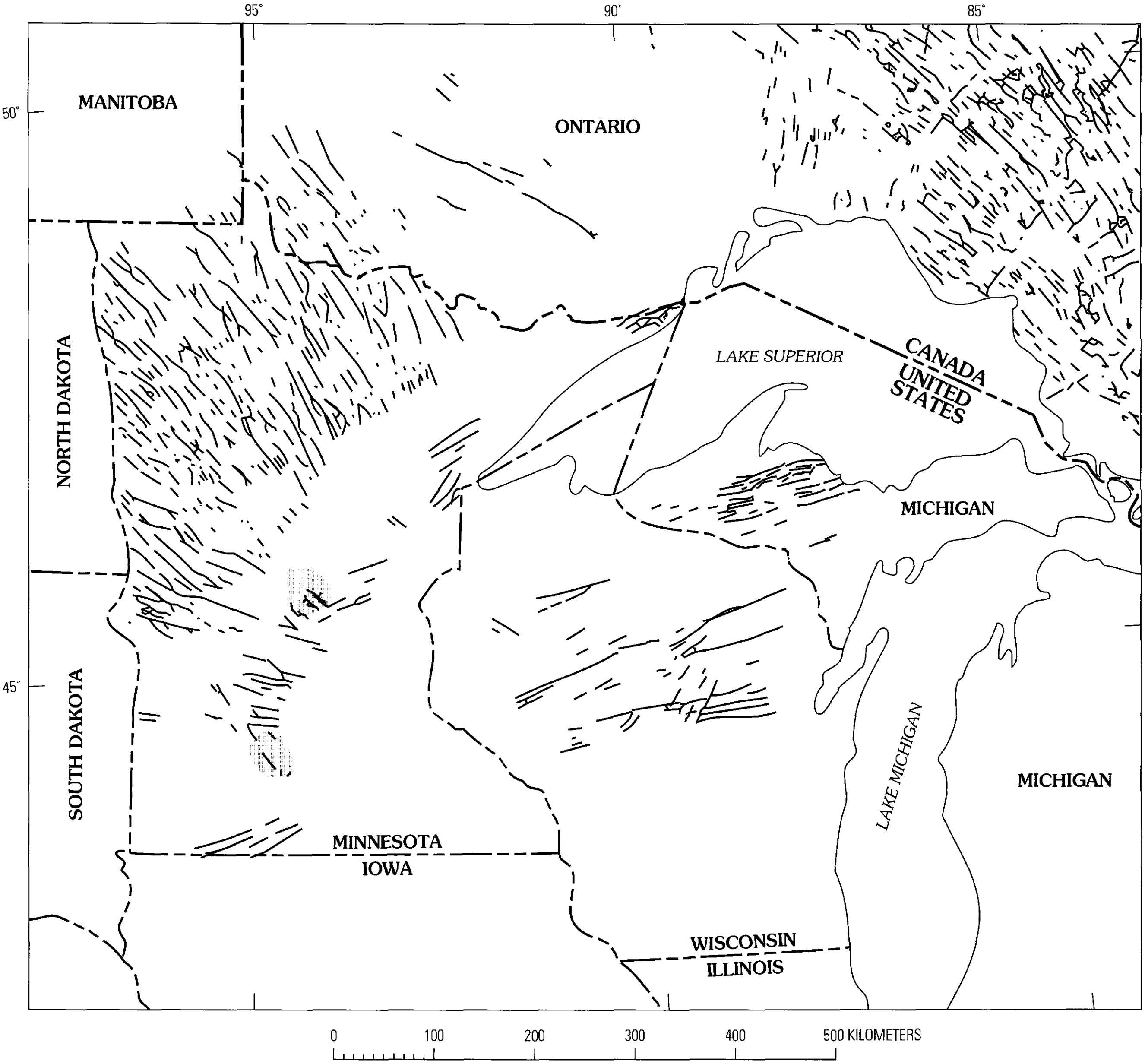

Figure 47. Dike swarms in Lake Superior region, mostly determined from aeromagnetic data. Shaded areas in Minnesota, areas containing numerous plugs of mafic and ultramafic rocks. Dikes in Canada from Fahrig and West (1986).

hinterland of the Animikie basin (Southwick and Halls, 1987).

Aeromagnetic data indicate that the Archean and Early Proterozoic rocks of central and southern Minnesota are cut by two dike swarms, one striking west-northwest and the other striking east-northeast (figs. 43, 47). The west-northwest-striking swarm intersects the western part of the Kenora-Kabetogama dike swarm in west-central Minnesota and Penokean orogen rocks in east-central Minnesota; the negative magnetic signatures associated with some of the dikes indicate reversed polarity. Two large dikes associated with this swarm are exposed in the Minnesota River Valley; these dikes have not been dated, but they may correlate with a nearby stock of gabbroic and granitic rocks which has been dated at 1,750 Ma (Goldich and others, 1961; Hanson, 1968). Dikes of the east-northeast-striking swarm cut across rocks of the Penokean orogen in east-central Minnesota and the Sioux Quartzite of southwestern Minnesota. The age of the east-northeast-striking dikes is not known, but is bracketed between 1,770 and 1,100 Ma (Southwick and Chandler, 1989).

Mafic dike swarms associated with the 1,100 Ma Midcontinent rift system are evident in the aeromagnetic data surrounding Lake Superior. In northeastern Minnesota (figs.

43, 47), a series of northeast-striking magnetic anomalies near the south end of the Duluth Complex correlates with 1- to 70-m-wide dikes of the Carlton County swarm (Green and others, 1987), and negative signatures indicate that most dikes have reversed polarity. Northeast-striking positive anomalies extending from the north end of the Duluth Complex (figs. 43, 44, and 47) reflect 100- to 500-m-wide dikes of the normally polarized Pigeon River swarm (Green and others, 1987). East-west negative anomalies south of Lake Superior (fig. 44) delineate 1- to 185-m-wide dikes of the reversely polarized Baraga-Marquette swarm (Green and others, 1987; King, 1990). East-northeast-striking anomalies that are predominantly negative in central Wisconsin (figs. 45, 47) may reflect a southern extension of the Baraga-Marquette swarm (King, 1990). On the basis of position and similarity in strike, the swarm in central Wisconsin may correlate with the east-northeast-striking swarm in central and southwestern Minnesota (fig. 47).

The high-resolution aeromagnetic data in Minnesota have also been useful in delineating mafic to ultramafic plugs in east-central and southwestern Minnesota (fig. 47), which produce swarms of small, circular highs of a few tens to a few hundreds of nanoteslas amplitude (a few are visible in fig. 43). A drill hole into one of the plugs in east-central Minnesota (fig. 47) encountered a mica-bearing olivine pyroxenite of possible lamprolite-kimberlite affinity (Southwick and Chandler, 1987). Magnetic model studies by Southwick and Chandler of this and surrounding plugs indicated near-vertical, pipelike bodies that extended to depths of at least 1.5–3.1 km. A few of the small circular anomalies in southwestern Minnesota (fig. 47) correspond to exposed plugs of gabbroic and granitic rocks in the Minnesota River Valley that have been dated at 1,750 Ma (Goldich and others, 1961; Hanson, 1968).

CONCLUDING REMARKS

Gravity and magnetic studies have been a critical component of Precambrian geologic investigations in the Lake Superior region during the last several years. One of the principal uses of these studies has been in geologic mapping, which is hampered by poor outcrop control in many areas. The improved geologic mapping has led to several major reinterpretations of the geology and tectonic evolution of the area. Quantitative analysis of gravity and magnetic data, including grid processing and a variety of modeling schemes, not only is an aid to geologic mapping but also provides valuable structural information in the third dimension. Model studies of gravity and magnetic data have been particularly useful supplements to seismic-reflection studies of the Midcontinent rift system.

Despite the recent progress attained in gravity and magnetic studies, much work remains to be done in the Lake Superior region. The scientific potential of new data sets produced through recent surveys and compilations is still largely untapped. All these new data sets are digitally based, and are amenable to the wide spectrum of quantitative analyses that are becoming available on inexpensive personal computers and work stations. Acquisition of new data is also needed, especially in areas of poor coverage in parts of Wisconsin and the Dakotas. All things considered, gravity and magnetic studies should continue to contribute significantly in the future to geologic investigations in the Lake Superior region.

ACKNOWLEDGMENTS

The Minnesota Legislature supported most of the high-resolution aeromagnetic surveying and much of the gravity surveying, as recommended by Legislative Commission on Minnesota Resources. This paper was prepared with support from the State Special Appropriation of the Minnesota Geological Survey. Dennis Teskey of the Geological Survey of Canada provided the shaded relief map of the GLIMPCE aeromagnetic data of Lake Superior, and Elizabeth King of the U.S. Geological Survey provided a shaded relief map of the northern Wisconsin aeromagnetic data.

METALLOGENY

By G.B. Morey *and* P.K. Sims

Precambrian rocks of the United States segment of the Lake Superior region historically have been a major national source of metallic ores. Iron-formation of both Archean and Early Proterozoic age has been the principal commodity, the total production having exceeded 5.1 billion metric tons of iron ore. Copper also has been an important commodity, more than 5.6 million metric tons of ore having been produced from Middle Proterozoic volcanic and sedimentary rocks on the Keweenaw Peninsula in Michigan. Identified but as yet undeveloped resources include volcanic-hosted massive sulfide deposits in Early Proterozoic rocks of the Wisconsin magmatic terranes, sedimentary rock-hosted manganese deposits in Early Proterozoic rocks of the Penokean orogen in east-central Minnesota, and gabbro-hosted copper-nickel deposits in the Middle Proterozoic Duluth Complex of northeastern Minnesota. More speculative resources include gold and base-metal sulfides in Archean greenstones and platinum-group elements in parts of the Duluth Complex (table 4). General references to the metallogeny of the region include Sims (1987), Sims and

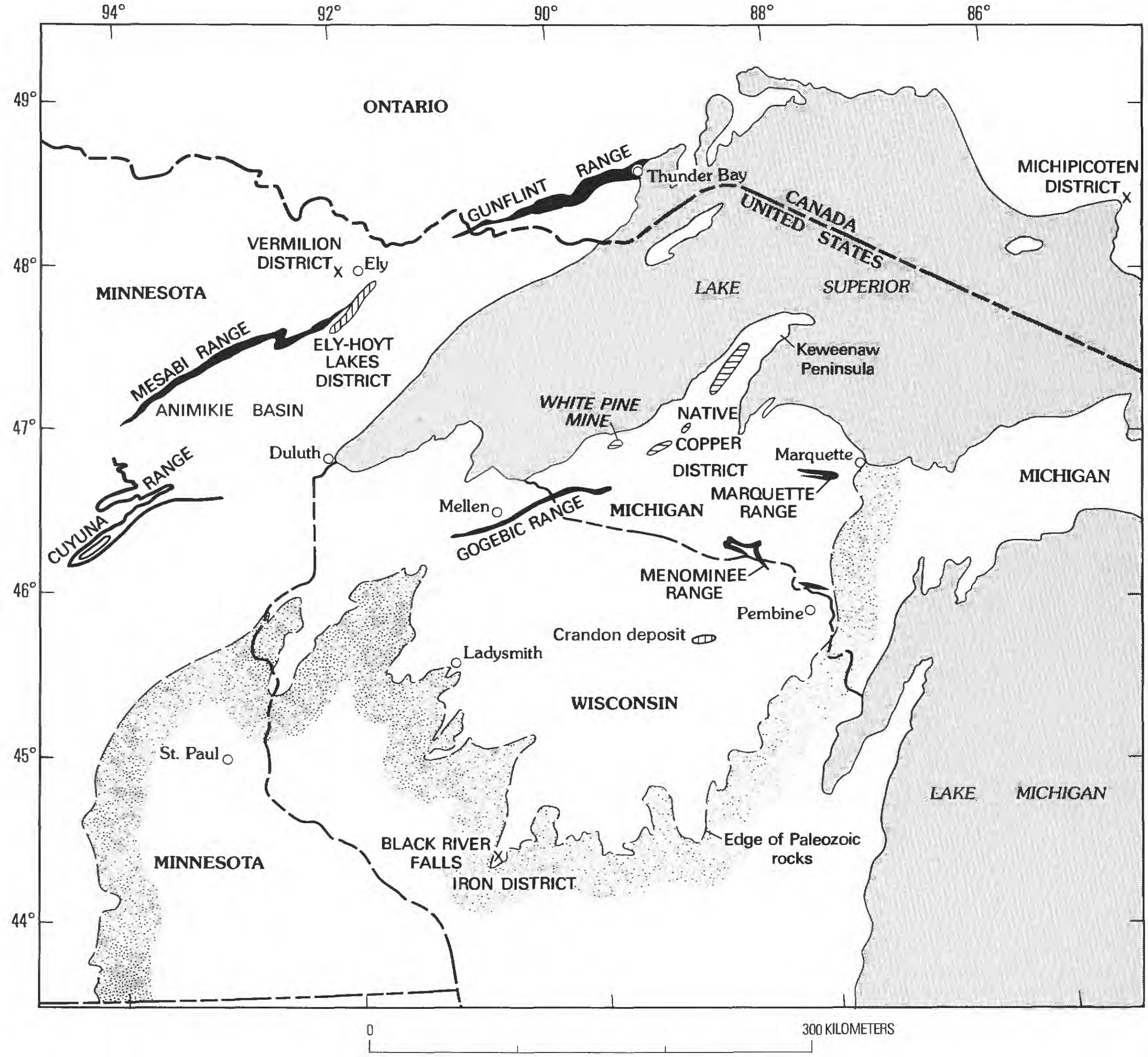

Figure 48. Major known and potential metallic mineral districts of Lake Superior region. Horizontal line pattern, known mining district; vertical line pattern, potential mining district; solid black, iron-formation. From Sims, 1976, 1987.

Table 4. Major Precambrian crust-forming events and associated major metalliferous deposits in the Lake Superior region.

Age (Ga)	Terrane	Tectonic setting	Description	Known mineral deposits	Probable and speculative resources
3.6–2.6	Minnesota River Valley subprovince.	Uncertain	Amphibolite-bearing migmatitic gneisses, younger gneisses and schists of greenstone affinity, and 2.6-Ga granite.	None reported	Volcanic-hosted massive-sulfide deposits. Volcanic-hosted gold deposits.
2.7–2.6	Wawa and adjacent subprovinces.	Oceanic arc	Tholeiitic basalt, basalt, komatiite, calc-alkaline basaltic andesite, andesite, and dacite; distinctly bimodal. Tonalite to granite intrusive rocks. Sutured to gneiss terrane in Late Archean.	Algoma-type iron-formation.	Volcanic-hosted massive-sulfide deposits. Volcanic-hosted gold deposits. Gabbro-hosted nickel-copper deposits.
2.1–≈1.9	Early Proterozoic epicratonic sedimentary rocks of Penokean orogen.	Shelf and foredeep environments on passive margin.	Bimodal basalt-rhyolite volcanic rocks and intercalated clastic and chemical sedimentary rocks.	Lake Superior-type iron-formation. Carbonate-hosted stratiform copper deposits.	None reported.
1.89–1.83	Early Proterozoic Wisconsin magmatic terranes of Penokean orogen.	Oceanic arc	Tholeiitic basalt and calc-alkaline andesite to rhyolite extrusive rocks and calc-alkaline gneiss and tonalite to granite. Sutured to continental margin ≈1.85 Ga.	Volcanic-hosted massive-sulfide deposits.	Volcanic-hosted gold deposits.
1.76	Early Proterozoic rhyolite-granite terrane of southern Wisconsin.	Extension	Anorogenic rhyolite and cogenetic epizonal granite; locally overlain by fluvial quartzite of the Baraboo interval.	None reported	Olympic Dam–type iron-copper-uranium-gold deposits.
1.2–1.0	Middle Proterozoic Midcontinent rift system.	Extension (intracontinental rift).	Aluminum-rich olivine tholeiite, high-iron tholeiite and rhyolite; tholeiitic intrusions.	Volcanic- and sedimentary-hosted copper deposits. Gabbro-hosted copper-nickel-cobalt deposits.	Redbed copper deposits. Gabbro-hosted PGE deposits.

others (1987), Morey (1989a), and Mudrey and Kalliokoski (1993). Figure 48 locates the major identified and potential metallic mineral districts in Precambrian rocks of the region.

DEPOSITS IN ARCHEAN ROCKS

Algoma-type iron-formation, as defined by Gross (1965), in the Vermilion district of the Archean Wawa subprovince, northern Minnesota (fig. 48), has been a significant source of natural iron ore (Sims, 1987). Production from 1884 until closure in 1967 totaled more than 89.2 million t (metric tons) of direct shipping ore and 4.8 million t of gravity concentrate. The district contains lenses of iron-formation intercalated on all scales with volcanic rocks, both mafic flows and felsic pyroclastic deposits, and with epiclastic sedimentary rocks, some of which are carbonaceous. Producing mines, however, were developed only in the world-famous Soudan Iron-formation near Tower-Soudan and in a smaller unnamed body of iron-formation at Ely. The iron ore deposits of both mines consisted principally of hard, massive hematite in several small, steeply dipping elongate lenses. Average composition ranged from 57 to 66 percent Fe, 2 to 8 percent SiO_2, 0.05 to 0.25 percent P_2O_5, and 0.4 to 2.0 percent Al_2O_3. Several ore bodies contained locally abundant native copper, chalcopyrite, and other copper minerals.

The iron-formations of the Vermilion district are exhalative deposits that most likely accumulated during lulls in volcanism and sedimentation. The largest iron-formation of the district, the Soudan Iron-formation, apparently formed during a prolonged hiatus after the completion of one volcanic cycle and before the start of the next. Today there is a general consensus that the high-grade hematite deposits formed by post-metamorphic hydrothermal processes in Late Archean time (Morey, 1983b).

Recent geophysical investigations have identified many units of Algoma-type iron-formation in northern Minnesota. Several of these appear to persist for many kilometers along strike. None, however, appear to have any substantial economic value.

Small quantities of iron-ore concentrate (less than 95 million t) were developed by the Jackson County Iron Company in Wisconsin. The production came from the Archean iron-formation at the Seven Mile Mound deposit east of Black River Falls in south-central Wisconsin (Chan and others, 1991). The ore body was a lensoid mass, some 900 m long and 550 m wide. It is intercalated within quartzofeldspathic schist and gneiss, generally metamorphosed to the upper greenschist to lower amphibolite facies. As such the iron-formation contains a metamorphic mineral assemblage including cummingtonite-grunerite, ferroactinolite, calcium-rich hornblende, garnet, and biotite, as well as appreciable magnetite and quartz.

Several small occurrences of gold were reported from Archean rocks in northern Minnesota as early as 1865 (Sims, 1972c), but only the Little American mine, on the Rainy Lake–Seine River fault on the Minnesota-Ontario boundary, produced any ore. About $4,600 worth of gold was extracted in 1894–1895 from a composite vein 1–2 m wide in sheared chloritic and biotitic schist. The vein also contained ankerite, pyrite, minor chalcopyrite, and tourmaline.

More than 20 gold prospects have been identified at one time or another in the Ishpeming greenstone belt in northern Michigan, but only the Ropes gold mine, about 25 km west of Marquette, has been a significant producer. About 6,500 kg of gold, together with a substantial amount of silver, was produced from 1882 to closing in 1985. The deposit still contains an appreciable resource estimated at 2.8 million t assaying 3.24 g gold per metric ton and minor silver (Brozdowski, 1989). The ore bodies occur in veins structurally concentrated in ductile shear zones (Brozdowski and others, 1986; Brozdowski, 1989). Higher grade deposits, which were mined in the past, are associated with quartz-tetrahedrite veins, whereas lower grade material mined in the 1980's occurs as disseminated grains associated with pyrite in the host rock.

Gold deposits in both the Little American and Ropes mines have attributes of the lode-gold deposits typically found in adjacent parts of Ontario north and northwest of Lake Superior. Most of this class of gold deposits occurs in Archean greenstone belts in the Wawa and Wabigoon subprovinces, where hundreds of mines have recorded productions ranging from less than 1.02 t to approximately 1,016 t of gold.

A consensus now exists that the productive lode-gold deposits in the Superior province of Canada are tabular, subvertical structures associated spatially with steeply dipping planar shear zones of brittle to ductile deformation that in turn are related to regional faults or regional zones of ductile shear. These transcurrent and oblique slip-shear features formed in Late Archean time, some 15–20 m.y. after regional deformation (Colvine, 1989, and references therein). The major structures are east-trending dextral transcurrent shear zones and northeast-trending sinistral faults that branch from and occur between the dextral shears. Similar structures have been identified in the southwestward extensions of the Wawa and Wabigoon greenstone belts in Minnesota and Michigan (Southwick, this volume).

The occurrence of Archean lode-gold deposits in steeply dipping deformation zones is of considerable exploration significance, because such zones can readily be identified in many kinds of geophysical imagery. However, individual gold-bearing systems are much smaller in size. Typically, the thickness of an individual vein system is measured in meters, and its strike and dip dimensions are measured in tens or a few hundreds of meters. Furthermore, the source of mineralizing fluids is debatable. That many lode-gold deposits are essentially the same age as late-tectonic syenite and lamprophyre bodies which were emplaced preferentially along the regional deformation zones (Wyman and

Kerrich, 1988) may have genetic significance. The identification of several such bodies in northern Minnesota supports the suggestion of Sims and Day (1992) that large gold deposits comparable to those known in Canada could be found there. An especially favorable area is the western Vermilion district (Alminas and others, 1992).

Although exploration has focused on lode gold, the Archean greenstone belts in the Lake Superior region possibly contain another class of gold deposits, namely stratiform deposits. Most of these deposits occur within units of carbonate or silicate (Algoma-type) iron-formation or their metamorphic equivalents. Geological, mineralogical, and geochemical relationships imply that the gold, as well as other enriched constituents such as sulfur and arsenic, is exhalative in origin, and thus was essentially in place prior to folding and metamorphism. The occurrence of ore bodies in former open spaces, as along the crests of folds, implies that the ore components, together with vein quartz, were mobilized and redistributed over distances of meters or possibly tens of meters during subsequent folding and metamorphism.

Archean rocks in northern Minnesota also have been widely explored for volcanic-hosted massive sulfide deposits, especially since 1970, when a State-wide geologic map (Sims, 1970) was published that showed for the first time the existence of several "greenstone belts" like those in Canada that contain base-metal sulfides. Exploration has not identified any sulfide deposits of probable economic significance, but drill holes have penetrated thick intervals of massive, submassive, and disseminated pyrite and pyrrhotite at several places, and thin, apparently discontinuous zones of zinc and copper have been encountered locally in the Birchdale-Indus, Minn., area (Ojakangas and others, 1977) along the Canadian border and in the Vermilion district (Sims and Morey, 1974). Geochemical studies by Alminas and others (1992) have further substantiated anomalous copper and zinc values in parts of the Vermilion district. Exploration ceased in the late 1970's for economic rather than geologic considerations.

Much information is available regarding volcanic-hosted base-metal sulfides that should aid exploration in Archean greenstone belts. Most known Archean occurrences belong to a primitive type of zinc and copper sulfide deposits that have been described by Hutchinson (1980). These deposits are spatially associated with rhyolitic rocks of tholeiitic affinity (Lesher and others, 1986) that occur within discrete eruptive centers. The deposits themselves, however, are concentrated within layers of epiclastic rocks and chert that accumulated during breaks in volcanic activity. The inferred coupling of sulfides and tholeiitic host rocks has been used as an exploration tool in Canada where rare-earth-element (REE) signatures in the felsic rocks have been used to discriminate between potentially barren and mineralized intervals. Although felsic rocks of tholeiitic affinity have been reported in Minnesota only in the Rainy Lake area (Day, 1985), the technique should have wider applicability in the future.

An additional aspect of greenstone belt metallogeny in the Lake Superior region is the possible presence of copper-nickel sulfide deposits associated with ultramafic host rocks. In Canada, these small but relatively rich deposits are concentrated in bodies of magnesium-rich peridotite that were emplaced either as sills and dikes or as komatiitic lavas. The two types of occurrences are related in that the peridotitic magmas in the hypabyssal rocks probably represent the residue remaining after eruption of the komatiitic lavas. Both sills of peridotite and flows of generally komatiitic composition have been recognized in the Vermilion district (Schulz, 1974, 1982).

DEPOSITS IN EARLY PROTEROZOIC ROCKS

Both the Early Proterozoic continental-margin rocks and the Wisconsin magmatic terranes contain known metallic deposits (Sims, 1987). The somewhat younger (≈1,760 Ma) rhyolite-granite terrane of southern Wisconsin lacks known metallic deposits but has some potential as host for mineral deposits of Olympic Dam type (Sims, 1990f; Enaudi and Oreskes, 1990). Lastly, the red quartz arenites of southern Wisconsin and southwestern Minnesota have some potential for Witwatersrand-type gold deposits and unconformity-related uranium deposits (Southwick, Morey, and Mossler, 1986).

Early Proterozoic iron-formations have yielded nearly 4.5 billion t of iron ore, or 92 percent of the iron ore shipped from the Lake Superior region since 1848. Production has come from several separate districts, or iron ranges, within the region (fig. 48), but the most productive by far has been the Mesabi range, in northern Minnesota. This world-class district has yielded 3.4 billion t of iron ore, or 67 percent of the total production from the Lake Superior region since mining began in 1892.

Early Proterozoic iron-formations in general are mixtures of two fundamental types: (1) Granular iron-formation, typically thin to thick bedded and either consisting of silt- and sand-size oolites and intraclasts set in a finer grained groundmass of microcrystalline chert or less commonly consisting of iron silicates and iron carbonates. Magnetite typically occurs as octahedra rarely more than 0.1 mm in diameter. In many beds of granular iron-formation, the individual octahedra are clustered together to form regularly to irregularly arranged beds, laminae, or patchy to mottled aggregates. (2) Nongranular iron-formation, generally fine grained, finely laminated, and consisting mostly, but not exclusively, of iron silicates and iron carbonates. It typically lacks oolites and intraclasts and contains little free quartz. Consequently, the two types have somewhat different compositions. On the Mesabi range, the total iron content is about the same in both types, but because of the magnetite, there is more ferric iron in the granular strata than in the nongranular strata. Conversely, there is less silica and more magnesium and aluminum in nongranular strata than in the granular strata (Morey, 1992).

Much of the iron ore mined in the Lake Superior region is so-called natural ores, which are the products of secondary oxidation and leaching of original iron-formation. Original iron-bearing minerals were oxidized to the ferric state, and at virtually the same time, calcium, magnesium, and much of the silica were removed by leaching. Most commonly, magnetite was altered to martite, iron silicates to goethite±hematite and kaolinite, and siderite to hematite or goethite. Much of the hematite-rich ore material that was produced was of the so-called "direct-shipping" variety—material that went directly from mine to blast furnace. (Other hematite-rich ores require some beneficiation, mainly to remove free silica, either by washing or by heavy-media techniques.) Today, production from direct shipping ores is scant, and most of the iron ore production involves mechanically separating primary magnetite from thick intervals of granular iron-formation and concentrating it to produce taconite pellets that contain 60 or more percent iron.

Early Proterozoic iron-formations of the Emily district and the North Range Group of the Cuyuna range (Morey, this volume) also constitute an important manganese resource (Morey, 1990). The size of the resource is not well constrained, but the North Range Group contains at least 455 million t of manganiferous iron-formation containing from 2 to 10 weight percent manganese that is available to open-pit mining to a depth of 45 m (Lewis, 1951). Of that total, at least 170 million t has an average grade of 10.45 weight percent manganese (Beltrame and others, 1981). As such the North Range Group contains approximately 46 percent of the known manganese resources in the United States (Dorr and others, 1973, p. 394, table 77). Nearly all of the estimate is based on data obtained in and around oxidized and leached bodies of manganiferous iron ore. Little is known regarding the resource contained in unaltered manganiferous iron-formation. Until recently the protolith was believed to be sedimentary and the manganese to occur principally as a manganiferous carbonate. Now, however, the protolith is shown to be submarine-hydrothermal in origin and primary manganese oxides to occur as thin to very thick layers intercalated with hematite-rich intervals, which contain appreciable rhodonite and rhodochrosite, as well as other exotic minerals such as barium-rich feldspar and aegirine (Cleland and others, 1992; McSwiggen and others, 1992b; Morey and others, 1992b).

Iron-formation of the Animikie Group in the Emily district, at the far north end of the Cuyuna range (Morey, this volume), also contains an appreciable manganese resource, possibly in excess of 2 million t averaging 20 weight percent or more manganese (Morey and others, 1991). Abundant manganese occurs as stratiform lenses at two stratigraphic levels. The larger of the two has a possible strike length of 1,600 m and an average thickness of 15–20 m. The oxides occur as disseminated grains, mottles, lenses, pods, and pore-filling cement in granular iron-formation, and are epigenetic (Morey and Southwick, 1993).

It has never been economically feasible to recover manganese from the Cuyuna range. The ore-bearing zones are fairly small and buried beneath as much as 60 m of glacial materials; but the main problem involves the lack of metallurgical methods that can effectively separate manganese from the iron without expending large quantities of energy.

Other known resources in the Early Proterozoic continental margin rocks are a carbonate-hosted copper deposit near Marquette, Mich., in the Kona Dolomite, near the base of the Marquette Range Supergroup (Morey, this volume). The deposit is broken by numerous high-angle faults, but estimates suggest that it contains about 1.016 billion t of potential ore containing 0.3 percent copper (Kirkham, 1989). The copper occurs as disseminated chalcocite, bornite, and chalcopyrite within a quartz-arenitic interval intercalated within dolostone. The deposit probably formed in a sabkha-like environment (Taylor, 1974).

More speculative resources in rocks of the Early Proterozoic continental margin assemblage include potentially large graphite resources in the Baraga Group in Michigan and in the Animikie Group in Minnesota (McSwiggen and Morey, 1989). Several of the graphitic intervals in Minnesota contain from 1–2 weight percent to as much as 44 weight percent carbon over stratigraphic intervals as much as 150 m thick, and typically have anomalous gold (350 ppb), silver (6 ppm), copper (2,500 ppm), and zinc (3,300 ppm) values. Thus, the terrane has some potential for sediment-hosted, stratabound copper-lead-zinc deposits similar to those found in the Churchill province in Canada and in the Rammelsberg ore bodies of Devonian age in Germany.

The Early Proterozoic magmatic terranes in northern Wisconsin contain significant massive sulfide deposits (Mudrey and others, 1991). To 1993, eight deposits totaling nearly 1.016 million t have been discovered. All belong to the primitive type of zinc and copper sulfide deposits of Hutchinson (1980), and accordingly are generally similar to massive sulfide deposits in Archean greenstone belts of the Superior province. All the deposits discovered so far are in the Pembine-Wausau terrane of Sims (this volume). DeMatties (1989) recognized two mineralized districts in the western part of the terrane—the Ladysmith district centered on the Flambeau deposit and the Samo district centered on the Ritchie Creek prospect (DeMatties, 1990)—and a third mineralized district at the east end of the terrane, centered on the very large Crandon deposit, which is 175 km east of Ladysmith.

From study of the stratigraphic framework in the Ladysmith and Samo districts, DeMatties (1989) recognized three separate favorable geologic environments, each with its own kind of mineralization. These are (1) a main volcanic-arc sequence of dominantly mafic volcanic rocks;

(2) a back-arc basin sequence of dominantly volcanogenic and volcaniclastic sedimentary rocks that also includes intermediate to mafic volcanic units as discrete volcanic successions; and (3) eruptive volcanic centers or complexes of rhyolitic composition. Important units in the back-arc basin sequence include layers, generally less than 30 m thick, of chlorite- and mica-rich argillite, which commonly are graphitic or sulfide bearing. The Thornapple deposit north of Ladysmith is an example of a deposit associated with the first environment (Mudrey and others, 1991). The Ritchie Creek prospect in the Samo district contains examples of sulfide mineralization associated with back-arc and rhyolitic eruptive environments (DeMatties, 1989). The Flambeau deposit in the Ladysmith district is a particularly good example of sulfide mineralization associated with a rhyolitic volcanic center (May, 1977).

The Flambeau deposit, which currently (1993) is being mined, is a supergene-enriched zone some 90 m thick that occurs beneath a cap of Upper Cambrian sandstone (May, 1977). The ore body contains about 10.5 percent copper and smaller amounts of precious metals.

No stratigraphic model similar to that of DeMatties (1989) has been made for the eastern part of the Pembine-Wausau terrane, where the Crandon deposit is located. This deposit, discovered in 1976, is of world-class size. It consists of a syngenetic zinc-rich massive sulfide zone and an underlying epigenetic copper-rich stringer-sulfide zone. Both are enclosed within a sequence of pyrite-bearing tuff and argillite, which in turn is intercalated with pyroclastic rocks of andesitic to dacitic composition (May and Schmidt, 1982; Schmidt, 1991). Reserves of nearly 44 million t averaging 8.40 percent zinc, 0.73 percent lead, 0.60 percent copper, 66 g silver/t, and 0.37 g gold/t have been identified in the massive ore zone. An additional 24.8 million t of ore averaging 1.80 percent copper, 0.7 percent zinc, 0.03 percent lead, 11.73 g silver/t, and 0.34 g gold/t occurs within the stringer ore zone.

The discovery of two additional deposits in the Samo district was announced in 1990, but little is known about either of them. The Bend deposit contains two ore zones (Mudrey and others, 1991). A copper-rich zone some 390 m long, 9 m thick, and 365 m wide contains at least 1.73 million t averaging 2.73 percent copper and 1.92 g gold/t. An underlying gold-rich zone, 330 m long, as much as 26 m wide, and extending more than 600 m below the surface, contains at least 945,000 t averaging 0.09 percent copper and 5.73 g gold/t.

A significant thickness of ore-grade massive sulfide also has been intersected at the Lynne prospect in northern Wisconsin. The first hole at the prospect intersected 39 m of polymetallic-mineralized rock averaging 22.7 percent zinc, 0.64 percent copper, 2.95 percent lead, 35.66 g silver/t, and 3.77 g gold/t (Borland, 1990). A subsequent 39 drill holes identified a preliminary mine resource of 6.2 million t grading 7.14 percent zinc, 11.66 g copper/t, 1.89 percent lead, 105.96 g silver/t, and 0.45 g gold/t (Kennedy and others, 1991). Much of the mineralization occurred within a sequence of chemical sedimentary rocks, which thicken and coalesce toward the central part of the deposit. Lenses, lobes, and stratiform bodies of massive to semimassive sulfide are intercalated with conformable units of talc, chert, laminated to disrupted carbonate, and thin beds of volcaniclastic and epiclastic sedimentary rocks. The aggregate thickness of the sulfide-rich interval in the center of the deposit exceeds 99 m. Kennedy and others (1991) have suggested that the ore zone formed as a hydrothermal mound followed by an undetermined amount of replacement mineralization. The mineralized interval is hosted within a thick sequence of pyroclastic rocks of rhyolitic composition, together with less abundant flows of dacitic to andesitic composition, and volcaniclastic sandstone and siltstone.

DEPOSITS IN MIDDLE PROTEROZOIC ROCKS

Stratified rocks of the Middle Proterozoic Keweenawan Supergroup contain the volcanic-hosted (White, 1968) and sediment-hosted (Ensign and others, 1968) copper deposits of the Keweenaw Peninsula, Mich. A large copper resource of marginal grade also occurs in troctolitic to gabbroic rocks of the Duluth Complex in northern Minnesota (Weiblen, 1993); the mineralized rock includes appreciable quantities of nickel and cobalt, as well as trace amounts of gold. Non-economic to marginally economic vanadium, titanium, and platinum-group elements (PGE) also have been found in several different rock types and structural settings within the complex. All the mineral occurrences are related to rifting processes and associated mafic igneous activity directly involving mantle-derived materials.

Considerably more speculative is the possibility that sedimentary rocks of the Keweenawan Supergroup could host red-bed copper deposits (Sims, Kisvarsanyi, and Morey, 1988) such as those found, for example, in a variety of Precambrian, Permian, and Cretaceous red-bed sequences, all in the Trans-Pecos region of Texas (Price and others, 1985). Most of these known deposits are small, but the ores are interesting because they typically contain appreciable amounts of silver.

Keweenawan stratified rocks of the Keweenaw district in Michigan have been a major source of copper and silver since the middle of the 19th century. Native copper occurs primarily as open-space fillings and replacements in flow-tops and conglomerates. Mineralization was controlled both by primary porosity and permeability and by tectonic fracturing. Weege and Pollock and others (1972) estimated production of 5.6 million t of copper from 370 million t of ore from 1845 through 1968. Wilband (1978) estimated that 21 of the native copper mines still contained nearly 49 million t of rock bearing more than 0.5 percent copper. The sedimentary rock-hosted deposits consist dominantly of disseminated chalcocite in dark laminated shale containing abundant pyrite and organic material, including petroleum. A deposit of economic size and grade developed in the Nonesuch Shale at the White Pine mine produced 1.38 million t of copper, having an average grade of 1.2 percent Cu, and 283,000 kg of silver between 1961 and 1980. Seasor and Brown (1986) estimated that the White Pine property still has reserves of 184 million t grading 1.1 percent copper and 6.72 g silver/t. A second noneconomic sedimentary-hosted deposit occurs some 20 km west of White Pine (Ensign and others, 1968), where 829,000 t of material containing 1.3 percent copper has been delineated (Wilband, 1978).

The distinctive mineralogical and zoning patterns and large copper:iron ratios of the Keweenawan copper deposits are characteristics of the "rift-related stratiform copper deposits" of Sawkins (1984). It has been generally accepted that the Keweenawan copper deposits were precipitated from heated copper-bearing solutions that formed in deeper parts of the rift system where the rocks were compacted and metamorphosed (White, 1968, 1971). In the volcanic-hosted deposits, the ascending solutions deposited most of the native copper in open spaces close to the boundary between the epidote and pumpellyite metamorphic zones. In the sedimentary rock-hosted deposits, the ascending solutions led to the replacement of pyrite by chalcocite (Ensign and others, 1968; Brown, 1971), probably shortly after deposition of the host rocks. The source of the copper has never been established with certainty, but Weiblen and others (1978) and Morey and Weiblen (1978) have proposed that the copper-charged solutions were produced by the diagenetic and low-grade metamorphic breakdown of copper-bearing basaltic detritus in the sedimentary sequences rather than by leaching of the basalts themselves.

Copper- and nickel-bearing sulfides in the Duluth Complex include chalcopyrite, cubanite, pentlandite, and pyrrhotite. Much of the sulfide-mineralized material is scattered throughout several troctolitic-gabbroic sequences emplaced in the lower part of the complex. Mineralized material has been found at only six localities and all material is of subeconomic grade; resource calculations for these localities indicate the presence of 4.1 billion t of material containing more than 0.5 percent copper. That resource has an average copper grade of 0.70 percent and a copper:nickel ratio of 3.33:1 (Listerud and Meineke, 1977).

Naldrett (1981) has assigned the copper-nickel deposits of the Duluth Complex to his class characterized by "intrusion-feeding flood-basalt activity associated with intracontinental rift zones," and has suggested that a magma derived directly from a mantle source was the source of the metals. However, a consensus now obtains that Early Proterozoic country rocks were the source of the sulfur (Weiblen and Morey, 1976; Mainwaring and Naldrett, 1977; Ripley, 1981).

Virtually all the world's platinum-group-element (PGE) production comes from deposits in layered mafic igneous complexes of Precambrian age, and the Duluth Complex, though not a perfect analog to PGE-producing complexes, warrants consideration as a potential PGE producer.

Background platinum content of unmineralized mafic and ultramafic rocks in general is approximately 10 ppb, with a range from 0.1 ppb to 500 ppb. To be of economic grade, a mean platinum value of 5 to 10 ppm is needed—an enrichment of three or four orders of magnitude. Concentrations of PGE in the Duluth Complex are associated with both copper-nickel sulfides and chromite-rich units. For example, Morton and Hauck (1989) reported anomalous PGE values of 1,000–3,000 ppb combined Pt+Pd in rocks with Cu:Cu+S ratios greater than 0.4. Similarly, Weiblen (1989), Dahlberg (1989), and Morton and Hauck (1989) have noted that several chrome spinel-bearing intervals in troctolitic rocks contain as much as 7,000 ppb combined Pt+Pd. Ryan and Weiblen (1984), Sabelin (1985), and Sabelin and others (1986) have shown that the PGE's occur as discrete phases—as native elements, as alloys of several native elements, and as several kinds of arsenides and sulfarsenides. That the PGE were introduced into the Duluth Complex as a minor orthomagmatic constituent is generally agreed; however, considerable disagreement (Morey, 1989b) still exists as to whether the PGE's were concentrated by normal magmatic processes or by the fluxing of hydrothermal (deuteric) PGE-enriched fluids through a partly crystallized magma. Regardless, the recent discoveries have created considerable exploration interest in the Duluth Complex.

UNCONVENTIONAL DEPOSITS

Mineral deposits that differ significantly from productive deposits in mineralogy, grade, or mode by which they may be exploited are termed unconventional (Barton, 1983). Such deposits are surprises, discovered by accident in forms or regions where they were not previously thought to exist. Mudrey and Kalliokoski (1993) have pointed out that a variety of unconventional mineral deposits could exist in the Lake Superior region, because its geologic history is so diverse and so poorly known in detail. Already recognized unconventional deposits include (1) supergene gold and base-metal sulfides in Archean greenstone belts in west-central and northwestern Minnesota; (2) kimberlites of post-Ordovician age in Michigan and possibly elsewhere in the Lake Superior region; and (3) petroleum in sedimentary rocks of the Midcontinent rift system. Mudrey and Kalliokoski (1993) have discussed several other possible mineral occurrences of the unconventional category.

Much of the Precambrian crust in the Lake Superior region has been affected by two periods of prolonged weathering that produced thick clay-rich saprolites of generally kaolinitic composition. The first of these weathering events occurred in Late Proterozoic to Late Cambrian time and probably affected much of the Lake Superior region. However, the resulting weathering profile is preserved at only a few places in Wisconsin (Chan and others, 1991) and in east-central Minnesota (Morey, 1972), generally beneath a cap of Upper Cambrian sandstone. Nonetheless, this weathering event was responsible for at least some of the oxidized and leached iron ore deposits in Early Proterozoic iron-formation in Michigan (Morey, 1983b) and the supergene sulfide deposits at the Flambeau mine near Ladysmith, Wis.

The second major weathering event probably took place during Late Cretaceous time. Evidence for it is widespread in the west half of Minnesota (Parham, 1970) in the form of weathering profiles on a variety of Precambrian rocks. In particular, this event was responsible for the hematite-rich natural iron ore deposits of the Mesabi range and at least some of the high-grade ores of the Cuyuna range (Morey, 1983b). Although weathering profiles characterized by a kaolinite-rich saprolite a few tens of meters thick are best preserved beneath a protective cover of Upper Cretaceous strata (Setterholm and others, 1989; Morey and Setterholm, in press), partly preserved profiles consisting of partly weathered saprock have been penetrated at many places at depths from 100 to 200 m where the Cretaceous cover has been stripped away and where the profiles are buried by a Quaternary cover. Kaolinite produced during this weathering event and reworked by subsequent sedimentary processes during Late Cretaceous time may be a significant economic commodity in the future.

Cannon and Mudrey's (1981) recognition of a kimberlite intrusion near Lake Ellen in northern Michigan provided a plausible explanation for the source of historical diamond finds in Pleistocene glacial deposits in the Lake Superior region. So far at least six kimberlite intrusions like the Lake Ellen body have been studied in northern Michigan (Jarvis and Kalliokoski, 1988). On geologic grounds they are middle Paleozoic or younger, as shown by xenoliths of Ordovician (Cannon and Mudrey, 1981) or Devonian (Jarvis and Kalliokoski, 1988) limestone. On the basis of some preliminary isotopic evidence, Jarvis and Kalliokoski (1988) concluded that all these kimberlites are Mesozoic, possibly Jurassic, in age, and, on the basis of petrologic evidence, that they are unlikely to be economic diamond deposits. However, these kimberlites were not the specific source for the large diamond finds in Wisconsin, and thus the potential for diamonds in other kimberlites in the area remains strong.

Detailed aeromagnetic studies in east-central Minnesota have identified a diffuse swarm of at least 70 small, sharp, subcircular, positive magnetic anomalies within an area of about 6,500 km^2. Rock recovered by core drilling from one anomaly proved to be a mantle-derived mica-bearing olivine pyroxenite of possible lamproite-kimberlite affinity (Southwick and Chandler, 1987). In addition, small intrusions of peridotite, first described by Lund (1950), which crop out at several places along the Minnesota River Valley near Franklin, were examined in detail by Chan (1990). She showed that they are chiefly composed of globules of orthopyroxene as large as 2 or 3 cm set in a serpentinized groundmass of olivine, pyroxene, spinel, garnet, and phlogopite. These peridotites have petrologic attributes comparable to those associated with mantle-derived peridotite elsewhere. Chan (in Chan and Weiblen, 1988) also identified an altered tuff in the Franklin area that contains several mineral phases having ultramafic compositions and typical volcanic textures. This ultramafic tuff most likely represents volcanic ejecta associated with the forceful emplacement of the peridotitic intrusions.

For several years, an extraordinary search for petroleum has been underway within sedimentary rocks of the Midcontinent rift system. The exploration effort stems in part from the fact that the Nonesuch Shale at the White Pine mine in Michigan contains solid and liquid hydrocarbons that have chemical attributes similar to those found in some crude oils currently being produced from strata of Paleozoic age in the Eastern United States. The oil is indigenous to the Nonesuch, which yields a Rb-Sr age of 1,052±5 Ma (Chaudhuri and Faure, 1967). The host rocks are cut by veins of calcite having included oil in small vugs. Calcite from one such vein yielded a Rb-Sr age of 1,047±35 Ma (Ruiz and others, 1984).

The fact that the oil in the Nonesuch Shale is definitely Precambrian in age has implications for petroleum exploration elsewhere in the rift system. The presence of an organic source rock in the White Pine area lends credibility to the expectation of organic-rich units in other parts of the rift system. A scenario for potentially successful exploration combines sedimentary sequences containing organic-rich units with tectonic environments predicting the presence of structural and stratigraphic traps; that indications of both these factors exist in the rift system implies that both sources and reservoirs may be amply developed (see Dickas, 1986).

Lastly, a study by Kelly and Nishioka (1985) has established that copper and hydrocarbons are similarly distributed in the Nonesuch Shale. As has been documented from many other places, organic-rich units serve as reductants and thus as traps for copper carried in oxidized, sulfate-rich ground waters. Thus metal explorationists as well as petroleum explorationists could benefit from the continuing search for hydrocarbons in the rift system.

FUTURE OUTLOOK

The Lake Superior region will continue to be a major source of iron ore. In 1992 the United States produced an estimated 56 million t of iron ore and iron-ore concentrate.

That value reflects 6.6 percent of total world production, estimated to be 844 million t. Of the total U.S. production in 1992, 41.5 million t or 74 percent was shipped from the Mesabi range. Similarly, more than 270 billion t of crude iron ore or 36 billion t of iron-ore concentrate is recoverable from Minnesota, Michigan, and Wisconsin with current open-pit mining methods (Marsden 1978a, 1978b; Cannon and others, 1978). Of that total, 48 billion t of ore grade or 8.8 billion t of concentrate occurs on the Mesabi range alone. Thus, the Mesabi range contains approximately 24 percent of the iron-ore concentrate in the Lake Superior region.

The volcanic- and sediment-hosted copper deposits of Middle Proterozoic age on the Keweenaw Peninsula also can be expected to continue to be a small but steady source of this metal. Even though the identified copper-nickel deposits in the Duluth Complex are marginal in grade in today's market (Sims, 1991b), they could become an important future source for copper as well as nickel and cobalt.

The base-metal sulfide deposits in the Early Proterozoic Pembine-Wausau terrane in northern Wisconsin also contain an appreciable copper resource, but more importantly they constitute a significant future source for zinc. According to Mudrey and others (1991), the zinc deposits in Wisconsin comprise 15 percent of the total zinc resource in the United States. The Crandon deposit alone, if it were to be brought on-line at the production levels originally proposed, could account for 25 percent of the total zinc production in the United States.

Although several discoveries have been made in the United States segment of the Lake Superior region, especially in Wisconsin, in the past decade, exploration is not easy. Much of the region is covered by glacial deposits and lacks appreciable bedrock outcrop. Still other parts have abundant cultural features that interfere with airborne geophysical surveys. Nonetheless, major discoveries undoubtedly remain for exploration geologists who have persistence, skill, faith, and luck. However, discovery is only the beginning. As soon as a resource is identified as a useful commodity and intended for exploration, development is subject to market forces and government regulation that will influence its ultimate availability.

REGIONAL TECTONIC ELEMENTS

By P.K. Sims

Several aspects of the geology of the Precambrian rocks along the south margin of the North America craton (Hoffman, 1989), including stratigraphic correlations, have been discussed and debated during this century. The purpose of this section is to examine some of the more important regional aspects in the light of our current knowledge. In a previous section, Morey discussed the long-debated stratigraphic correlations of Early Proterozoic rocks between the Lake Superior and Lake Huron regions.

ARCHEAN CRATONS

The two Archean cratons in north-central United States (pl. 2), which are separated by the Early Proterozoic Trans-Hudson orogen, differ with respect to lithotypes, structure, and geophysical character. These data together with Sm-Nd data (Peterman and Futa, 1988) clearly demonstrate that the Superior province and the Wyoming province are not correlative as might be expected by simple lateral extension across the Trans-Hudson orogen (Sims, Peterman, and others, 1991). Like much of the Superior province, the westward basement extension in the eastern Dakotas (as determined from drill cores) is juvenile Late Archean crust with Sm-Nd model ages of 2.76 to 2.67 Ga. In contrast, most of the Wyoming province is composed of crust that was extracted from the mantle in Early and Middle Archean time (3.6–3.1 Ga). Late Archean (2.7 Ga) granitic plutons, for the most part, also carry a Sm-Nd signature of this older continental crust.

The southernmost (United States) part of the Superior province consists of an Early-Late Archean gneiss terrane (Minnesota River Valley subprovince; pl. 1; Southwick, this volume), to the south, and Late Archean dominantly arc-related terranes to the north. From north to south, the arc terranes are designated the Wabigoon, Quetico, and Wawa subprovinces (Southwick, this volume; Card, 1990; Card and King, 1992). The Great Lakes tectonic zone separates the gneiss terrane from the arc-related terranes (Sims, this volume). Sims and others (1981) have suggested that the eastern extension of the GLTZ is the Murray fault system (pl. 2), a north-verging reverse fault or thrust fault on the north shore of Lake Superior that dips about 70° S. (Zolnai and others, 1984). If so, the Huron Supergroup rocks to the south of the Murray fault system should be underlain by Archean gneiss.

The Wyoming province is both smaller and less well exposed than the Superior province. Archean rocks crop out only in the cores of Laramide ranges in Wyoming and southwestern Montana (Hoffman, 1989). The south margin of the province is exposed in southeastern Wyoming in the Sierra Madre and Medicine Bow Mountains (Houston and Karlstrom, 1992; Houston and others, 1992), and possibly in the Richeau Hills (Houston, 1992), on the east side of the Laramie Range. In contrast to the United States segment of the Superior province, high-grade ortho- and paragneisses predominate, and greenstone belts are generally small and sparse. It is not known whether the province has been coherent since 3.6 to 3.1 Ga, the age of Nd crust-formation, or represents a tectonic collage assembled about 2.9 to 2.6 Ga.

EARLY PROTEROZOIC COLLISIONAL OROGENS

Three Early Proterozoic collisional orogens have been delineated along the south margin of the North American craton (pl. 2): (1) the Penokean, which extends from northeast Nebraska through the Lake Superior and Lake Huron regions to the Middle Proterozoic Grenville tectonic zone; (2) the Trans-Hudson, which separates and welds the Superior and Wyoming cratons; and (3) the Central Plains, which truncates the two Archean cratons and the Trans-Hudson and Penokean orogens, to the south.

PENOKEAN OROGEN

The Penokean orogen consists of a deformed, south-facing, passive continental margin prism overlying Archean basement of the Superior province that is juxtaposed on the south with accreted magmatic arc terranes termed the Wisconsin magmatic terranes (Sims and others, 1989). In the Lake Superior region, the continental margin assemblage consists of a lower rifted passive margin sequence (rift and platform deposits) overstepped northward by a synorogenic foredeep sequence (Southwick and others, 1988; Barovich and others, 1989). A thick-skinned foreland fold-and-thrust belt north of the Niagara fault zone (Klasner and Sims, 1993) passes cratonward (northward) into autochthonous Early Proterozoic sedimentary rocks (Gregg, 1993). The sedimentary prism, assigned to the Marquette Range Supergroup (Michigan and Wisconsin) and the Animikie, Mille Lacs, and North Range Groups (Minnesota), was deposited in the approximate interval 2.2–1.9 Ga.

In northern Michigan and Wisconsin, the suture (Niagara fault zone) that juxtaposes Wisconsin magmatic terranes with the continental margin trends easterly and overrides the deformed continental margin; suturing occurred at about 1.85 Ga (Sims and others, 1989). In Minnesota, to the west of the Midcontinent rift (pl. 2), the presumed equivalent (Malmo discontinuity; Southwick and others, 1988) of the Niagara fault zone makes a conspicuous oroclinal bend to the south and is inferred to continue southward through northwest Iowa as the Spirit Lake trend (Anderson and Black, 1983). The Penokean orogen is truncated in the subsurface in northeast Nebraska by the Central Plains orogen (pl. 2; Sims and Peterman, 1986; Sims, Peterman, and others, 1991).

East of the east arm of the Midcontinent rift (pl. 2), the continental margin sequence (Huron Supergroup) is overlain locally by foredeep deposits (Whitewater Group) of the Penokean orogen, as discussed by Morey (this volume). The Huron Supergroup is a south-facing sequence (see Sims and others, 1981, and references therein) consisting of four unconformity-bounded, northerly tapering, onlapping clastic wedges; the sequence was deposited during Early Proterozoic crustal stretching (and ocean basin formation?) along the south margin of the Superior province. The Huron Supergroup formed in the interval 2,500–2,200 Ma, and thus is distinctly older than the Marquette Range Supergroup. Deformation of rocks of the Huron Supergroup occurred before 2,200 Ma (age of the Nipissing Diabase) and again during the approximate interval 1.9–1.85 Ga (Zolnai and others, 1984). The younger deformation was most intense; it compressed the edge of the Superior craton and its cover of Huron deposits as thrusting propagated northward and eventually deformed the 1,850 Ma Sudbury Nickel Irruptive. Structures that formed in the thicker (southward) part of the clastic wedge during ductile deformation were rotated during the thrusting to their present near-vertical position. The cause of the pre-2,200 Ma deformation is enigmatic. The younger, more intense deformation (1,900–1,850 Ma), however, has been attributed to the Penokean orogeny (Zolnai and others, 1984). Accordingly, volcanic arc complexes equivalent to the Wisconsin magmatic terranes possibly underlie Lake Huron to the south of the exposed rocks of the Huron Supergroup. That the older, pre-2,200 Ma deformation is related to continent-arc collision is unlikely inasmuch as the oldest known arc-related rocks in the Wisconsin magmatic terranes are about 1,890 Ma (Sims and others, 1989), at least 300 m.y. younger than the pre-2,200 Ma event. The relative antiquity of the Huron deposits and the documented multiple deformation of the rocks (Zolnai and others, 1984) indicate that the Huron segment of the Superior margin probably is a compound passive continental margin. Presumably, Huron Supergroup deposits formed on an older margin prior to 2,200 Ma, then were deformed pre-2,200 Ma; subsequently these rocks, including the Whitewater Group in the Sudbury area, as well as the subjacent Archean basement were deformed again by the north-verging Penokean collision.

TRANS-HUDSON OROGEN

The Trans-Hudson orogen has been traced southward in the subsurface from exposures in northern Saskatchewan and Manitoba (Hoffman, 1989, and references therein), mainly by magnetic anomaly maps (Green and others, 1985). The orogen is characterized by dominantly north trending magnetic anomalies that reflect a north-trending structural grain. Isotopic age data indicate that the main igneous and metamorphic events occurred between 1,910 and 1,800 Ma (Sims, Peterman, and others, 1991).

The outline of the Trans-Hudson orogen as shown on plate 2 is inferred from aeromagnetic and gravity data and sparse drill holes, as well as extrapolation from exposures in northern Canada. An eastern unit (unit XWb; Superior-Churchill boundary zone; Sims, Peterman, and others, 1991), correlated with the Thompson belt in Canada, consists of probable reworked Archean basement and a dominantly metasedimentary cover. A central magnetic region (unit Xvc), to the west, consists dominantly of moderately dense rocks, such as intermediate-mafic volcanic rocks, probably mainly arc derived rocks, at least locally metamorphosed to granulite grade. Another magnetic belt, the western magnetic region (unit Xvw; Sims, Peterman, and others, 1991), composes the western part of the orogen. It appears to consist mainly of gneisses or metasedimentary rocks. Sims, Peterman, and others (1991) included the Black Hills uplift of Laramide age in the Trans-Hudson orogen because of its north-northwest-trending structural grain, typical of that elsewhere in the orogen. It is inferred to lie within a wedge-shaped, probably fault bounded block in western South Dakota that consists of Early Proterozoic metasedimentary and metavolcanic rocks that overlie reworked Archean basement rocks. It has been tentatively correlated (Sims, Peterman, and others, 1991) with the Cree Lake zone in northern Saskatchewan, as described by Lewry and others (1985).

A COCORP (Consortium for Continental Reflection Profiling) deep seismic-reflection survey across the northern part of the Trans-Hudson orogen and adjacent cratons (beneath the Phanerozoic Williston basin) at about lat 48°30' N. provides a transect across the orogen (Nelson and others, 1993). At about long 101° W., deep reflections dip eastward, beneath rocks of the Superior-Churchill boundary zone (compare Sims, Peterman, and others, 1991). This suggests that rocks of the Central magnetic region have been subducted beneath Archean rocks of the Superior province. At about long 103° W., mid- to lower crustal rocks reflect a broad antiform beneath the center of the Williston basin. Near the west margin of the orogen, as mapped by Sims, Peterman, and others (1991), a thick band of west-dipping reflectors (of western magnetic region) which may mark a Proterozoic subduction zone truncates a thick band of subhorizontal reflectors to the west, presumably within the Wyoming craton.

The seismic-reflection data suggest that both the east and west margins of the Trans-Hudson orogen are subducted beneath respective Archean basement rocks. In the Manitoba-Saskatchewan segment of the orogen, on the contrary, Hoffman (1989) and others have interpreted a southeastern foreland belt (Thompson belt) as having been thrust southeastward over the Superior province. The northwest margin (hinterland) is interpreted as involving major thrust stacking (mainly south- to southeast-vergent) attributed to accretion of island arcs and sedimentary rocks due to north- to northwest subduction, followed by oblique sinistral collision between the Hearne and Superior cratons.

CENTRAL PLAINS OROGEN

The Central Plains orogen (pl. 2; Sims and Peterman, 1986; Bickford and others, 1986) is the eastern, subsurface extension of the Early Proterozoic fold belt (Colorado province; Reed and others, 1987) exposed in the Laramide basement uplifts of Colorado and southeastern Wyoming. It is an arcuate belt (convex to the north) of metamorphic and igneous rocks formed during the approximate time span 1,800–1,630 Ma that truncates from west to east the Archean Wyoming craton, the Early Proterozoic Trans-Hudson orogen, the Archean Superior craton, and the Penokean orogen (pl. 2). Its northern boundary, the Cheyenne belt (Houston and others, 1979), is exposed in the Medicine Bow and Sierra Madre Mountains (Houston and Karlstrom, 1992), and is projected on the basis of geology and geophysics northeastward through the Laramie Range into southern South Dakota and northeastern Nebraska (Sims, Peterman, and others, 1991). In eastern Nebraska, the Central Plains orogen apparently is truncated by the Middle Proterozoic Midcontinent rift.

The Cheyenne belt, as exposed in the Medicine Bow Mountains, is a major shear zone a maximum of 5 km wide that separates Archean gneiss and Early Proterozoic miogeoclinal rocks of the Wyoming craton to the north from 1.8–1.63 Ga arc-related gneisses and calc-alkalic plutons to the south (Houston and others, 1989). In the Medicine Bow Mountains, the shear zone consists of four major mylonite zones that merge westward into a single ductile shear zone (Duebendorfer and Houston, 1987). The northern mylonite zone of the Cheyenne belt separates Early Proterozoic metasedimentary rocks of the upper part of the Libby Creek Group, to the north, from gneisses of the Bear Lake block to the south. The southern shear zone, the Rambler, defines the north margin of the volcanic arc complex to the south (Karlstrom and Houston, 1984). Neodymium data (Ball and Farmer, 1991) indicate that the rocks in the Cheyenne belt within the Medicine Bow Mountains are part of the Wyoming craton, inasmuch as they have εNd values and Archean model ages similar to those of the Libby Creek Group, north of the northern mylonite zone. The Nd data indicate that rocks in both the lower and upper parts of the Libby Creek Group were derived from Archean crust. The data require that the Rambler shear zone is the suture between the Archean continental margin and the Early Proterozoic arc rocks. Earlier, Duebendorfer and Houston (1987) suggested that the northern mylonite zone, also called the Mullen Creek–Nash Fork shear zone (Houston and McCallum, 1961), is the suture between the two terranes. Further, the Nd data indicate that foredeep sedimentary rocks with 2.4 to 2.0 Ga Nd model ages are absent north of the Cheyenne belt, contrary to the earlier suggestion of Houston (1992) that the French Slate, at the top of the upper part of Libby Creek Group (Houston and others, 1992) is a probable foredeep deposit.

Geologic mapping by Houston and Karlstrom (1992) and others in the Medicine Bow Mountains indicates that both the lower and upper parts of the Libby Creek Group are fault (thrust) bounded and have been thrust northward over the Early Proterozoic Deep Lake Group and underlying Archean rocks. Accordingly, these rocks are presumably allochthonous and constitute a tectonic stratigraphy.

PROBABLE COMPOUND NATURE OF SOUTH MARGIN OF WYOMING PROVINCE

Sims, Peterman, and others (1991) suggested that the east margin of the Wyoming province is marked by major faults that coincide with steep gravity gradients and locally with conspicuous magnetic contrasts. One of these faults (Hartville-Rawhide fault) is exposed in the Hartville uplift (Snyder, 1980). South of the Hartville uplift (pl. 2), it merges with, or more likely, converges toward the Cheyenne belt, as shown on plate 2. I propose that this fault extends westward and is either the Reservoir Lake fault or the Lewis Lake fault (Houston and Karlstrom, 1992) in the Medicine Bow Mountains. The Reservoir Lake fault is 6 km north of the Cheyenne belt (northern mylonite zone) and thrusts lower Libby Creek Group rocks northward over the Deep Lake Group and its subjacent Archean basement. This interpretation is supported by contrasting structures in the Deep Lake and Libby Creek Groups (Houston and Karlstrom, 1992). These authors concluded that the deformation of the Deep Lake Group took place between about 2.5 and 2.1 Ga (pamphlet, p. 16). With this interpretation, the Deep Lake Group was deformed by a collision, possibly along the Reservoir Lake fault, before 2.1 Ga. Following this collision, rifting of the continental margin resulted in deposition of the lower Libby Creek Group, and later this margin evolved into an open marine basin in which rocks of the upper Libby Creek Group were deposited (Karlstrom and others, 1983).

Encroachment and, eventually, collision of the arc rocks of the Central Plains orogen at ≈1.75 Ga deformed the continental-margin Libby Creek Group rocks in a fold-and-thrust belt north of the Cheyenne belt. The complex nature of this collision is reflected by the apparent Archean rocks within the Cheyenne belt itself, as indicated by Nd data (Ball and Farmer, 1991).

REGIONAL CORRELATIONS OF EARLY PROTEROZOIC MARGIN SEQUENCES

The rough delineation of buried portions of the Early Proterozoic orogens and Archean cratons in north-central United States (pl. 2) provides a tectonic framework within which proposed stratigraphic correlations can be examined.

Correlation of the thick siliciclastics (Snowy Pass Supergroup) in the Medicine Bow Mountains (fig. 49) and Sierra Madre with the Huron Supergroup has been proposed by Young (1973) and Houston and others (1992, fig. 21). Although this correlation is appealing because of lithologic similarities and a similar order of succession, the assemblages are homotaxial because they formed on different cratonic margins and are not necessarily contemporaneous. Also, Houston and others (1992) have proposed, on the basis of lithotype similarities, that the upper part of the Early Proterozoic Libby Creek Group may correlate with shelf deposits of the Marquette Range Supergroup in Michigan. However, these assemblages too are homotaxial because

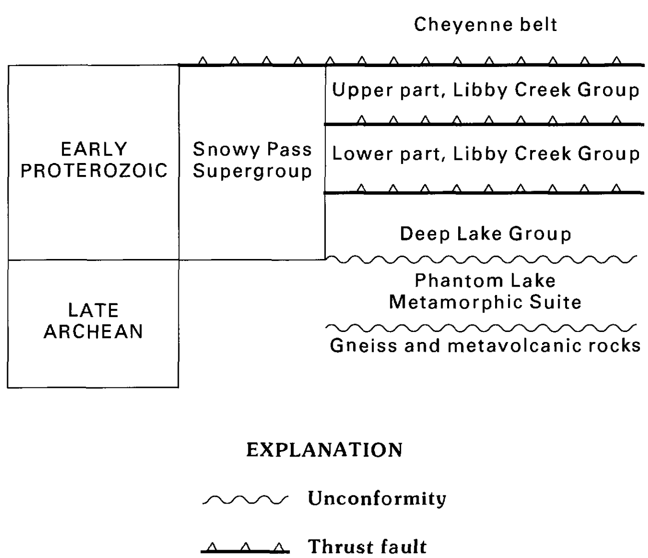

Figure 49. Stratigraphic succession of metasedimentary rocks of the Medicine Bow Mountains, southeastern Wyoming. Wavy line, unconformity; sawteeth, upper plate of thrust fault. Modified from Houston and others, 1992.

they formed on different cratonic margins, and they probably are not contemporaneous. Further, Houston's (1992) assumption that the youngest units (Towner Greenstone and French Slate) of the Libby Creek Group are foredeep deposits is not supported by subsequent Nd data. Neodymium studies by Ball and Farmer (1991) indicate that these rocks were derived from the Wyoming Archean craton, and not from erosion of advancing thrust sheets of the encroaching arc terrane of the Central Plains orogen.

Although contemporaneity of the dominantly siliciclastic sequences of Early Proterozoic age in the Lake Huron region and southeast Wyoming is improbable, the lithologic similarities can be explained by their tectonic settings. As discussed by Morey (this volume), the stratigraphic sequences can be equated generally with Holocene sequences along passive margins, as well as with older Phanerozoic sequences such as appear in the Appalachians. The dominantly siliciclastic sequences of Early Proterozoic age can be explained by erosion of the quartz-rich Archean rocks in the source areas. These eroded materials accumulated along the rifted continental margins. Why analogs to the foredeep deposits in the Lake Superior region are apparently lacking in southeastern Wyoming is not known, but perhaps such deposits were formed and then subsequently eroded, as suggested by Ball and Farmer (1991).

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